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THREE CRITICAL ENVIRONMENTS OF DEPOSITION, AND CRITERIA FOR RECOGNITION OF ROCKS DEPOSITED IN EACH OF THEM

BY JOHN LYON RICH

ABSTRACT

In any body of standing water subject to wave action, three depositional environments can be distinguished. They are here named the *unda*, the *clino*, and the *fondo* environments. In each, conditions are distinctive, but within each relatively uniform. The physical agencies operative in each of them, together with the kind and rate of sedimentation and rate of subsidence, control sedimentary facies and produce distinctive features of composition and bedding which are criteria for determining the depositional environments.

Wave base, the greatest depth to which the bottom is stirred by waves during storms, is a critical factor. Above it, on the *undaform*, the sediments are relatively coarse, generally fossiliferous, rippled and cross-bedded, and show a peculiar "wavy" type of relatively thin bedding. The *clinoform* is the sloping surface extending from wave base down to the generally flat floor of the water body, here called the *fondoform*. In the clino environment—*i.e.*, that on the clinoform—mud is dominant. Fossils, except for microforms in the later sediments, tend to be scarce. The bedding shows a distinctive alternation of thin and remarkably even layers of silt- and clay-sized material. Flute markings, probably produced during storms by density currents, and various indications of slumping and flowage of and in the sediments are common. The fondo environment is characterized by fine material with even, uniform, and generally rather massive bedding, but uniform, thin-bedded alternations of silt and shale without flute markings may develop close to the base of the adjoining clinoform.

Changes in sea level shift these environments and produce characteristic sequences of strata. One inferred effect of a drop in sea level is the deposition, all along the outer edge of the undaform, of a belt of sands—here called undaform-edge sands—capable of becoming reservoirs for oil and gas. Such sand zones are likely to be misinterpreted as "shoreline trends."

Rocks deposited in the unda environment are relatively coarse, being most commonly sandstone, coarse siltstone, fragmental limestone, or oölite. The bedding is distinct and commonly has a characteristic waviness. Ripple marks, cross-bedding, flow-and-plunge structure, and lenticularity are characteristic.

Criteria for clino rocks are: texture fine enough for the material to have been carried in suspension, bedding relatively thin and remarkably even, and the presence of flow markings and flutings and of evidences of gravity sliding.

Criteria for fondo rocks are: fine grain, lack of secondary sorting, generally massive bedding, and a composition generally derived from clays and oozes.

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Introduction and Acknowledgments

On the floor of any body of standing water in which appreciable wave action occurs, three physiographic units or environments can be recognized, in each of which the physical and biological conditions are distinctive. These, in order outward from the shore, are: (1) that part of the floor which lies in the zone of wave action and in which, therefore, the bottom is repeatedly stirred and reworked by storm waves; this the writer proposes to call the *unda* environment, from the Latin word for wave. (2) the sloping part of the floor which extends from wave base down to the more or less level deeper parts; this the writer calls the *clino* environment. (3) all of the floor not compassed by the other two; for it the term *fondo* (Spanish for bottom) environment is suggested.

The sedimentary facies formed in these environments are the *unda* facies, the *clino* facies, and the *fondo* facies. One might also speak of unda deposits, unda beds, etc. For rock units formed in each environment the terms *undathem*, *clinothem*, and *fondothem* are proposed. For the topographic units represented by each, the term *undaform* is proposed for the subaqueous land form produced by the erosive and constructive action of the waves during the development of the subaqueous profile of equilibrium; and the term *clinoform* for the subaqueous land form analogous to the well-known "continental slope" of the oceans or to the foreset beds of a delta. For the subaqueous land form constituting the main floor of the water body, the word *fondoform* is proposed.

After a long period of stationary sea level during which the marine subaqueous profile of equilibrium (Fenneman, 1902) had been brought to full maturity, the undaform would correspond with the continental shelf, the clinoform with the continental slope, and the fondoform with the abyssal ocean bottom beyond the base of the slope. But the foregoing are only particular examples of the application of the terms proposed. Those terms must apply to small water bodies as well as to large, and to interior seas or lakes as well as to the ocean. In the special case of the continental shelf, they must be applicable to deposits of clino or fondo types made on the floor of the former shelf after a rise in sea level has placed parts of it below wave base.

These three environments are the most important keys to sedimentary facies because in them material of three different lithologic types is being deposited simultaneously under three sets of conditions, each of which has its own particular profound influence on living organisms.

Long ago, Chamberlin (1913; 1914a; 1914b) recognized that the sediments of the three environments are distinct, and he described each type briefly. Since then most geologists have tended to overlook the fact that rocks formed in each environment must retain features of composition and bedding so distinctive that when the criteria are adequately worked out it should be possible to examine any sedimentary rock and determine with a high degree of certainty in which environment it was deposited.

An important purpose of this paper is to make clear that a particular rock does not ordinarily indicate the total composition of the material being transported to the water body at the time of its deposition, but rather, it represents one of three very different segregations of that material that are deposited simultaneously, one in each of the three environments.

The terms "neritic", "bathyal", and "abyssal" are not employed here, because they connote specific depth ranges that do not correspond with the actual limits of the physical environments under discussion.

Wave base is a critical plane separating the unda from the clino environment. Its depth depends on the size of the water body concerned and on the intensity of the storms to which it

is exposed. It may range from 10 feet or less for a small lake to probably as much as 600 feet for places exposed to the full sweep of a stormy ocean.

The profile of a mature undaform is the profile of equilibrium for wave action. It drops rather steeply for a short distance out from shore, then slopes down very gently across the undaform to a point close to its outer edge where, in many instances, it becomes convex upward for a short distance as it makes the transition into the greater steepness of the clinoform (Fig. 1).

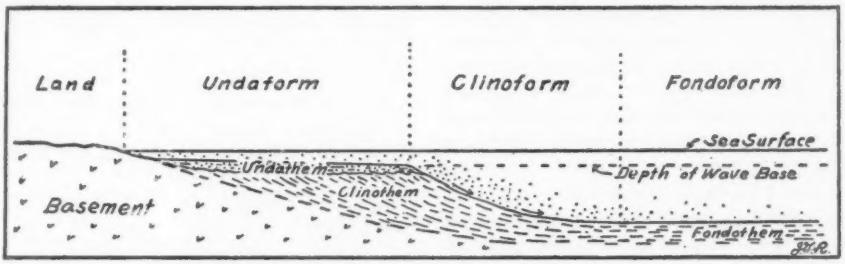

FIGURE 1.—SKETCH ILLUSTRATING DEFINITIONS

Undaform, clinoform, fondoform, undathem, clinothem, fondothem, and wave base and the distribution of muddy water after a storm.
Muddy water shown by stippling; density currents by arrows.
Vertical scale greatly exaggerated.

The clinoform is that inclined subaqueous depositional plane extending from the outer edge of the undaform down to the fondo level of the water body concerned. It normally starts with a short upward-convex curve at the top and ends with a concave curve at the bottom. The length of a clinoform depends on the vertical distance between wave base and the fondo level, and on its angle of inclination. It might connect an undaform with a fondo level only a few feet below it; or the difference in elevation between those two units might be many thousands of feet, as it is for the continental slope off the eastern coast of the United States.

The fondoform may have considerable irregularity, but, in general, its irregularities are being reduced by deposition. A submerged irregular topography or tectonic activity may produce considerable angles of slope in the fondo region that could cause sliding of the sediments and other features which might resemble those of the clino environment, but significant differences in the total assemblage of diagnostic features should ordinarily make it possible to distinguish fondo sediments, even if formed on an irregular bottom.

This paper is built on a foundation laid in a preceding paper (Rich, 1950) in which the features associated especially with the clino environment were described and analyzed. The interpretations there worked out are used in developing the criteria of environment here used. Many references to the literature used in working out those criteria will be found in the earlier paper.

When the earlier paper was written, the writer did not realize the necessity for introducing new terms in order to prevent confusion with common terms already used, sometimes loosely, in other senses. The words "shelf", "slope", and "bottom", with carefully defined limited connotations, were there used for undaform, clinoform, and fondoform, or, as adjectives.

This paper is primarily an attempt to develop criteria by which the principles and clues to interpretation worked out in the earlier paper can be applied to the practical task of interpreting environments of deposition from the physical and other features of sedimentary rocks.

The writer wishes to express his indebtedness to John Rodgers, Shepard W. Lowman, Kenneth E. Caster, and Gordon Rittenhouse, all of whom read the manuscript and contributed helpful suggestions.

Processes at Work in the Unda, Clino, and Fondo Environments

General Considerations

The discussion in this paper is necessarily in part deductive because, until recently, conditions on the floors of large bodies of water were little known, and even now are too imperfectly known for us to base our discussion solely on the results of direct observation. Where the statements here made rest on such observations they are made in positive form; where they are purely or mainly deductive they are expressed conditionally. The writer is a firm believer in deductive reasoning, both as a guide to thinking and as a guide to later field observations, and he has not hesitated to use it here.

The discussions are based on normal conditions and processes. No comprehensive attempt is made to explore the permutations and combinations that might result from such things as local tectonic disturbances, the immediate proximity of a large river mouth, or rapid fluctuations of sea level. In general, the effects of such departures from the normal should be readily deducible from the principles here developed.

Unda Environment

The most significant process of the unda environment is the repeated agitation of the water by waves and currents. Wave action frequently throws the finer material into suspension. Currents then readily move it, and ultimately (barring subsidence of the undaform below wave base before winnowing is complete) most of it must be carried out to settle permanently in deep water.

This gradual elimination of the finer particles causes a concentration of the coarser elements of the sediment load on the floor of the undaform. This coarser concentrate is shifted around on the bottom by currents, and, assuming a stable undaform, such of it as is not ground fine by the process or by wave agitation must ultimately be worked out to the edge and dumped into deeper water. If the quantity so dumped is large, it tends to overload the upper part of the clinoform and to produce instability that may cause gravity sliding.

Tidal currents on an undaform facing an open ocean must be important agents in redistributing the coarser material and in working it out toward the edge and into deeper water. Rough calculations indicate that, for a 3-foot tide on an undaform 100 miles wide, those currents, during periods of maximum flow, might reach a velocity as great as 3 inches per second. This, though by itself possibly insufficient for bottom traction, would assist greatly in transporting material lifted at intervals by the waves. (*See* Sverdrup, Johnson, and Fleming, 1942, p. 958–959.)

Undertow produced by on-shore winds (Fenneman, 1902) also tends always to move material toward the outer edge of the undaform. Probably such currents would not ordinarily be as swift as tidal currents, but, under favorable topographic conditions of the undaform surface, they might become effective for bottom traction, and always would assist in moving outward all fine material in suspension and any coarser material the waves temporarily raised from the bottom.

Any major ocean current which happens to flow over the undaform, even one not swift enough to lift material off the bottom, would also aid in moving material temporarily put into suspension by the waves. Currents would always be effective in causing a net movement of fine suspended sediment off the undaform into deeper water.

The cumulative effect of all these processes acting on a maturely graded undaform must be to eliminate the finer material gradually by working it out to deeper water; to leave behind a concentrate of the coarser sediment on the floor of the undaform; and, because of repeated reworking, to produce in this coarse material an irregular lenticularity of bedding, cross-bedding on a moderate scale, and ripple marks, presumably of both oscillation and current type.

On the continental shelves of the world (corresponding only in part to the definition of undaform here used) Shepard (1932) and Shepard and Cohee (1936) report that the prevailing sediments are sandy, and commonly

as coarse near the outer edge of the shelf as are those close to the shore. The coarser material of these shelves is sand and even gravel or, very commonly, broken shells (lime-sand or coquenite).

Exceptions to the prevailing coarseness of unda sediments occur where subsidence is relatively rapid in regions near the mouths of large rivers bringing in great quantities of sand, clay, and silt. Under such conditions the time available for reworking by waves on the undaform may not be great enough to permit the winnowing out and removal of all the fine material. Consequently the rocks, as preserved, may be composed mainly of fine sand and silt. Such a condition is reported (Shepard, 1949) for the northwestern part of the Gulf of Mexico and for the coastal regions of eastern Asia which are influenced by the great amounts of mud being carried into island-protected seas by the Yellow and the Yangtze rivers.

The writer believes that the sharpness of the break in slope at the edge of the undaform depends mainly on the coarseness of the material delivered by bottom traction to its edge —the coarser the material, the sharper the break—and on its quantity relative to that of fine material carried in suspension—the smaller the proportion of fine material, the sharper the break. Additional factors would be the velocity of currents carrying suspended material away from the edge of the undaform, and the velocity of tidal currents moving alternately in and out across its edge. Where tides are high and the undaform is exceptionally wide, these currents might be swift enough to rework, to a considerable depth below wave base, the material at the outer edge of the undaform and on the upper part of the clinoform and produce an ebb-and-flow type of small-scale cross-bedding which would be preserved as the outer edge of the undaform advanced into deeper water.[1] The effect of such a zone would probably be to decrease the sharpness of the edge of the undaform and to make the junction between undaform and clinoform convex upward.

On a stable, maturely graded undaform, little permanent deposition would occur, and the available sediment supply would be used mainly in building forward the undaform and clinoform units. But on a sinking undaform, subsiding slowly enough so that the sediment supplied always kept the surface of the undaform within the range of wave action, great thicknesses of sediment could accumulate, all having unda characteristics.

If sinking is somewhat more rapid than grading of the undaform to its profile of equilibrium, and if only a slight difference in depth exists between the surface of the undaform and that of the fondoform, a peculiar interbedding of unda and fondo sediments, such as is found in the Cincinnatian rocks of the Cincinnati region, might develop. There, calcareous shales of fondo type, in beds ranging from a few inches to 1 or 2 feet thick, are interstratified with unda shell-hash limestones 2–8 inches thick. It is suggested that while the Cincinnatian sediments were accumulating the ocean bottom lay slightly below the depth of wave base for ordinary storms, so that fondo muds could accumulate, but that during occasional exceptionally severe storms the zone of wave agitation extended to a greater depth, reworking a few feet of rather sparsely fossiliferous fondo muds, putting the fine material into suspension, and leaving on the bottom a residual mass of shell fragments. As the storm subsided, the material in suspension settled over the wave-sorted shell hash as a layer of unfossiliferous mud. Part or all of this layer would be preserved along with a cover of sparsely fossiliferous fondo mud if gradual sinking in the interval before the next great storm carried the bottom below the wave base for that storm. By such a process, thick accumulations of alternating fondo muds and unda fragmental limestones could be made.[2]

Sediments deposited in the unda environment near the mouth of a large river carrying great quantities of mud might be abnormally fine and poorly sorted if deposition accompanied sinking so rapid that insufficient time was given for wave and current winnowing of the

[1] A zone of such material is found beneath the presumable unda beds of the Berea sandstone in southeastern Ohio.

[2] The ideas expressed in the preceding paragraph were presented as a working hypothesis by Walter H. Bucher to a local field group about 15 years ago.

material on the undaform. Such a condition has been reported in the northwestern part of the Gulf of Mexico. Conditions there would be more favorable for it than if the water body were exposed to greater and more continuous strong waves.

Conditions such as those of the northwestern Gulf of Mexico should, however, be considered somewhat exceptional because of the great amount of mud being carried in by the Mississippi and the relative calmness of the Gulf except for occasional hurricanes. Also, the permanent unda sediments of the northwestern Gulf may not be so fine as reports have indicated, because the samples from an ordinary dredge, or short cores, might consist only of fine material that had settled temporarily after having been put into suspension by the latest great hurricane, rather than of the material which had finally come to permanent rest.

Conditions in the unda environment favor development of a luxuriant bottom-dwelling fauna of the larger shellfish, corals, and brachiopods. The unda zone lies entirely in depths through which light penetrates, and, except in the immediate vicinity of the mouth of a large muddy river, the water is clear except during severe storms because the bottom is composed mainly of the coarser sediments left behind as the finer material is put into suspension by the waves and carried away into deep water.

Clino Environment

General description.—The most distinctive features of the clino environment are the inclination of the surface of the clinoform and the freedom from wave-caused disturbances of the water. Secondary features are prevailing muddiness, great and oft-repeated variations in sediment supply, deposition dominantly from suspension, density currents periodically flowing down the slope, and gravity sliding and/or intra- or interstratal flowage, intensified probably by a tendency toward the development of quicksand in the coarser beds. These distinctive physical features of the clino environment impart readily recognizable characteristics to its sediments.

Under stable conditions, the undaform and clinoform are continually built outward. The coarse sediment is worked along the bottom to the edge of the undaform and there dumped onto the upper part of the clinoform. The finer material in suspension is carried out into deep water by the undertow and other currents. Thus coarse material tends to accumulate at the top of the clinoform and to oversteepen it; while the fine sediment in suspension tends to spread widely and to be carried down the slope of the clinoform by density currents to accumulate mainly near its foot and on the adjacent fondoform, thus decreasing the angle of slope of the clinoform. Preponderance of coarse sediment tends to produce steep slopes; preponderance of fine sediment, gentle slopes.

Oversteepening of the upper part of the clinoform by the coarse sediment may proceed until imbalance is sufficient to cause the overloaded part to slide bodily down the slope as a subaqueous landslip. Such slips may carry blocks of unda material far down the slope and deposit them in the midst of totally different clino or fondo sediments (Jones, 1937; Bailey, Collet, and Field, 1928; Kuenen and Migliorini, 1950).

Whether a clinoform is relatively steep at the top and consequently subject to subaqueous landslips, or has slopes too gentle for the more violent types of sliding is believed to depend mainly on the relative proportions of coarse to fine material supplied to it and on the strength of currents moving suspended material off the undaform into deeper water.

However, a clinoform built forward into a very deep and capacious water body would tend to develop steep slopes and to maintain them for a long time because deposition on the lower parts of the clinoform would be inhibited by powerful density currents induced by great length of clinoform slope, and shortening of that slope by filling of the basin would be very slow.

A combination of fine material, strong currents, and low available relief would be especially favorable for the production of gently sloping clinoforms.

The actual gradients of clino slopes should vary from that of the subaqueous angle of repose for gravel to those so low that the beds might not readily be recognized as having

been laid down in a clino environment, and so low as to preclude even the sliding of soft muds and the intrastratal flowage of "quick" silts.

On the steeper parts of the continental slope off the east coast of the United States (Veatch and Smith, 1939) the inclination of the upper 6000 feet of the clinoform ranges between 585 feet per mile (6° 20′) and a more common figure of about 333 feet per mile (3° 37′). At greater depths the inclination gradually decreases.

For the delta of the Nile, Barrell (1912, p. 388) gives a chart which shows a maximum inclination of 82 feet per mile (53′) for that part of the delta front between the depths of 50 and 200 meters; 220 feet per mile (2° 24′) for that between 200 m. and 1000 m.; and 176 feet per mile (1° 54′) for the average gradient from 200 m. to 2000 m.

Thus, the inclination of the Nile delta front is considerably less than the maximum for the continental slope off eastern United States.

Material of silt and clay sizes in suspension spreads far before complete settling. It produces very even and uniform bedding, and single beds persist for great distances with little variation in thickness. Bedding is graded, but, as a rule, not conspicuously. Grading is likely to be more a change from coarser to finer with distance from the edge of the undaform than from coarser to finer from the bottom to the top of a single bed.

Because deposition is from suspension, cross-bedding and ripple marks should be uncommon. They might, however, be produced by exceptionally strong density currents or, in some localities, by ordinary ocean currents. If produced by density currents, ripple marks should show movement directly down the slope of the clinoform.

As has been explained, the fine silt-and clay-sized material delivered to the undaform by rivers, together with that produced by attrition of the coarser materials on the undaform, is put into suspension during storms and carried out to the deeper water over the clinoform. After every great storm a considerable quantity of silt- and clay-sized material in suspension is carried off the undaform and settles widely over the clinoform and adjacent fondoform. Every great storm that roils the water on the undaform must therefore produce a single bed of silt on the clinoform. In general, the material of this bed should become finer in a direction away from the edge of the undaform and it also might thin in the same direction, but, because density currents tend to scour the upper part of the clinoform, such a bed may be thickest on the lower parts of the clinoform or on adjacent parts of the fondoform.

As each storm subsides, the finer constituents of the suspended load settle over the first-deposited coarser silts, and, during the periods between storms, only fine clay-sized material is laid down. In this way, accumulations are built up consisting of alternating beds of silt and clay or of lime-silt and calcareous ooze—later to become siltstones and shales, respectively, or their calcareous equivalents. These beds should be relatively thin because the material composing each can be assumed to have been deposited during a single storm.

Because of the inclination of the clino surface, sediments deposited in that environment are particularly subject to the effects of gravitative imbalance. This may be expressed by flow and minor sliding or creeping of the material within or between the bedding units—intrastratal flow (Pl. 2, fig. 1) or interstratal flow (Pl. 2, fig. 2) as well as by the large-scale landslips mentioned.

The prevailing fineness of the material deposited in the clino environment, plus the fact that the clinoform offers the first opportunity for the fine material to come permanently to rest, insures that the clino environment will be a muddy one and, under stable conditions, the locus of the most active permanent deposition. It would therefore be very unfavorable for any bottom-dwelling aquatic life except for certain mud-loving forms and some of the Foraminifera. Sediments accumulated there would generally be poorly fossiliferous, and the few fossils found would be either those of mud-dwellers or of pelagic forms that lived in the waters above.

Density currents and their effects.—A severe storm over the undaform must put into suspension a large amount of silt and clay. Since muddy water is heavier than clear water in proportion to the amount of sediment which it contains, sediment-laden water slides down the sloping surface of the clinoform as a density

current. Such a current apparently can attain considerable velocity (Daly, 1935; Kuenen, 1937; 1938; 1947; Kuenen and Migliorini, 1950). Its sediment load makes such a current decidedly abrasive, and it scours the surface of the clay deposited since the last great storm.

The writer (Rich, 1950) has shown that the peculiar striations and "flow markings" (Gefliessmarken) commonly found on the under sides of siltstone beds are casts of striations and flutings produced on the surface of mud by the scouring action of water flowing as a sheet over it. These markings also indicate the direction of flow when the flutings were made, and differences in the nature and intensity of the markings give considerable evidence as to the velocity of the currents. Certain groovings that not uncommonly accompany the flowage flutings and generally approximately parallel them, the writer thinks were made by water-logged woody materials and other solid objects dragged over the muds by the density currents.

As the storm subsides and the density current slows to a stop, abrasion ceases, and the silt in suspension settles to the bottom, forming the siltstone beds we observe.

The lower limit of scouring by the density current generated by a storm undoubtedly varies with the intensity of the current. It is farther down the slope for the more vigorous currents generated by the most severe storms. This may explain why the bottoms of some siltstone beds of a clino sequence show casts of flow markings and flutings produced by moving water, while others not only show none, but may even preserve undisturbed worm trails (Rich, 1950).

The above-described process can be observed from the campus of Cornell University at Ithaca, N. Y. Cayuga lake, whose south end lies about 1½ miles away from and 400 feet lower than that vantage point, is about 35 miles long and trends north-south. Whenever a strong north wind blows the waves stir the water over the subaqueous portion of the delta (undaform) that extends across the whole south end of the lake for about a mile out from the south shore. An observer on the campus sees a sharp, straight division line extending across the lake and separating the wave-stirred muddy water to the south from the normal clear blue water of the deeper parts of the lake to the north.

Inasmuch as a strong north wind must move surface water toward the south, an undertow must return the water northward beneath the surface. This undertow, as long as the high wind persists, must constantly carry mud-laden water off the undaform into deep water. That it should form a density current flowing down the front of the delta (the clinoform) is as inevitable as that muddy water is heavier than clear. That the density current is effective is shown by the sharpness and straightness of the line separating clear from muddy water. The silt so carried out must inevitably form a single definite, evenly distributed bed covering the delta front (clinoform) below wave base and the adjacent parts of the lake bottom (fondoform). As the storm subsides, a layer of finer material must be deposited over the bed formed during the storm, and this would be followed during the next severe storm by another coarser bed.

A large subaqueous landslip caused by oversteepening at the upper part of the clinoform may start a self-accelerating density current of high specific gravity capable of transporting blocks of considerable size, torn loose by the landslip, far out onto the lower parts of the clino or even into the fondo environment (Kuenen and Migliorini, 1950). Material suspended in such a density current, as it settles, produces graded bedding and, as Kuenen and Migliorini have shown, might produce the poor sorting of materials characteristic of graywacke.

Some of the peculiar conglomerates and breccias occasionally found in deep-water deposits are probably best explained by such high specific gravity density currents, whose effects must differ considerably from those of the ordinary density currents caused by fine material put into suspension on the undaform and carried out to deeper water as above described.

Role of compaction water.—The writer believes another effective process in the clino environment is the conversion of the coarser silt beds, after burial, into what are essentially fine-grained quicksands. The silt beds afford natural channels for the return, toward the surface, of water being squeezed by compaction out of the

finer muds interbedded on the clinoform with the silts. Any loose silt or sand through which water is moving upward becomes "quick" whenever conditions permit expansion of volume. This process seems to be common and is believed to be responsible for the general occurrence of intra- and interstratal flowage phenomena in sandstone or siltstone beds and not in or on the adjacent shales.

When a siltstone or sandstone bed on a clinoform becomes "quick", either differential flowage occurs within the bed without disturbing the strata above and below it (Pl. 2, fig. 1), or the whole bed is deformed, along with those immediately adjacent (Pl. 2, fig. 2; *see also* Rich, 1950). Deformation of the latter kind is common also in beds of unda type.

Fondo Environment

In the fondo environment the water is generally quiet, though locally and/or temporarily it may be moved by general currents and by waning density currents descending from the clinoform. The latter could affect only relatively small areas adjacent to the foot of the clinoform.

Deposition on the fondoform must be mainly from suspension, but bottom-dwelling organisms undoubtedly contribute a minor part. The sediment settling from suspension consists mainly of fine terrigenous material and of the remains of pelagic organisms living mostly in the surface water.

The sediment of the fondo environment is rarely, if ever, rearranged or sorted after it once reaches the bottom. Little reason is known for the development of thin and varied bedding except close to the base of the clinoform where the storm-generated showers of silt and clay would produce alternating bedding gradually grading into very thin laminations with increasing distance from the base of the clinoform. Preservation of the fine lamination of these beds depends on the amount of reworking by bottom life. Remote from the base of the clinoform, thick, massive bedding should be expected because of uniformity of sediment supply and of depositional conditions.

Whatever its thickness, fondo bedding is very even.

Remains of the larger bottom-dwelling organisms and of the larger forms of pelagic life, such as the shells of cephalopods or the teeth of sharks, are scattered through fondo deposits like plums in a pudding, entirely without sorting.

Deposition in the fondo environment must ordinarily be slow, and a long time would generally be represented by only a small thickness of sediment. Especially would this be true in the deeper waters far from land where, under certain conditions, most or all of the calcareous debris and calcareous remains of pelagic organisms would disappear by solution.

Fondo sediment may be deposited with abnormal rapidity in the lagoons of coral atolls and in other localities where wave action is slight, even in shallow water, and sediment supply from organic sources or from chemical precipitation is relatively abundant.

In inclosed basins, where the water is unaerated, toxic conditions would preclude the existence of bottom-dwelling scavengers and permit the preservation of noncalcareous organic material. Such basins would be favorable for the accumulation and preservation of such material because its finely divided condition and light weight would prevent its deposition in the agitated waters on the undaforms. Such light-weight organic material might settle in considerable quantity on the clinoforms, but there bottom-dwelling scavengers might destroy it, and such of it as escaped would be obscured by the much greater quantities of clay and silt settling in that environment. Consequently, the great accumulations of oil shales would originate in the toxic inclosed basins.

Criteria for Recognition of Sedimentary Rocks as of Unda, Clino, or Fondo Origin

Evaluation of Criteria

Recognition of environments of deposition from diagnostic features of rock composition, texture, and bedding is in part new, and not all the criteria have been thoroughly tested. Those here suggested undoubtedly can be sharpened, and additional ones will be discovered as rocks are more carefully studied with criteria for environmental origin in mind. Some of the criteria probably will need modification, but

the writer believes that from the physical and faunal features of a rock it will ultimately be possible to determine accurately the environment in which it was deposited.

The reader should remember that we are concerned exclusively with rocks originating in bodies of standing water and not with those formed from sediments deposited on the land by wind or running water.

Sedimentary Rocks of Unda Origin

A rock formed from sediments accumulated under conditions prevailing in the unda environment should have some or all of the following characteristics:

Texture—Relatively coarse, grains ranging in size from that of gravel through sand to coarse silt.

Composition—Conglomerate, sandstone, coarse siltstone, fragmental limestone ("shell-hash"), oölite, or coquenite.

Bedding—Moderately thin-bedded with a peculiar type of waviness illustrated in Figures 1 and 2 of Plate 1, but especially well shown in Figure 2 of Plate 1; cross-bedding; flow-and-plunge structure; individual beds not persistent, lenticular, and prevailingly cut out within short distances; bedding apparently even when viewed from a distance too great to reveal its small-scale irregularities; ripple marks common and of either oscillation or current type.

A rock showing all or most of the above features of texture, composition, and bedding can be confidently classed as of unda origin.

The only rocks not deposited in the unda environment, that might show some of these features would be the undaform-edge sands (defined on a following page) deposited at the upper edge of the clinoform or carried part way down it by density currents. Texture and composition of these might be the same as for unda rocks, but the bedding and associations would be significantly different. Specifically, all features of bedding produced by wave working would be absent, including the wavy type of bedding illustrated in Figure 2 of Plate 1, flow-and-plunge structures, cross-bedding, and erosional lenticularity. Coarse-textured rocks deposited on the upper part of the clinoform undoubtedly would be interbedded with shales, as are the siltstones characteristic of that en-

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PLATE 1.—UNDA AND CLINO BEDDING

FIGURE 1.—CROSS-BEDDING OF UNDA-TYPE LIMESTONE

Grand Tower limestone at Devil's Bake Oven near Grand Tower, Ill. Constituents are coarse sand-sized fragments of shells and crinoids.

FIGURE 2.—UNDA-TYPE BEDDING IN GRAND TOWER LIMESTONE AT DEVIL'S BAKE OVEN NEAR GRAND TOWER, ILL.

Note relative thinness of bedding, wavy top and bottom contacts, and generally uniform composition of the material.

FIGURE 3.—CLINO-TYPE BEDDING

Sea cliff at base of Constitution Hill, Aberystwyth, Wales. Note remarkable uniformity of bedding and small-scale alternations of siltstone (protruding beds) and dark shale. No individual bed of siltstone is as much as 1 foot thick. Many of the siltstone beds show internal contortions such as those of Figure 1 of Plate 2, and flute markings and striations such as those of Figure 1 of Plate 3.

PLATE 2.—INTRASTRATAL FLOWAGE FEATURES; INTERSTRATAL FLOWAGE FEATURES; FONDO BEDDING AND AUTO-BRECCIATION IN LIMESTONE

FIGURE 1.—INTRASTRATAL CONTORTIONS SUCH AS ARE COMMON IN SILTSTONE BEDS AT ABERYSTWYTH

This bed is 6 inches thick. Dark, slaty shale between siltstone beds. Knife length 3.3 inches.

FIGURE 2.—INTERSTRATAL FLOWAGE PHENOMENA IN FINE SANDSTONE

At approximately the Berea horizon north of Waverly, Ohio.

FIGURE 3.—LOOKING DOWN AT SURFACE OF FONDO-TYPE ROCKFORD LIMESTONE

Immediately above New Albany shale 1 mile west of Memphis, Indiana. Crinoid stems scattered without sorting in fine, dense lithographic-type limestone. Shows nontectonic brecciation, such as is common in some dense limestones, extending in irregular bands through the rock.

FIGURE 1

FIGURE 2

FIGURE 3

UNDA AND CLINO BEDDING

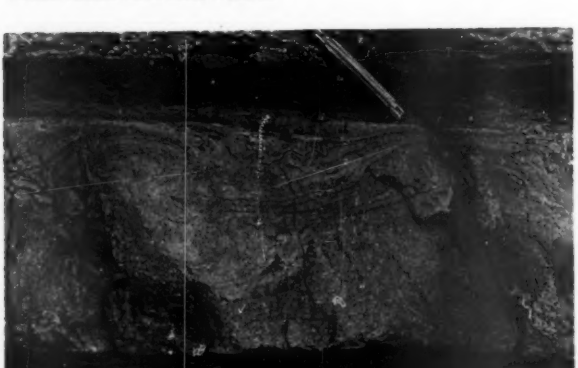

FIGURE 1

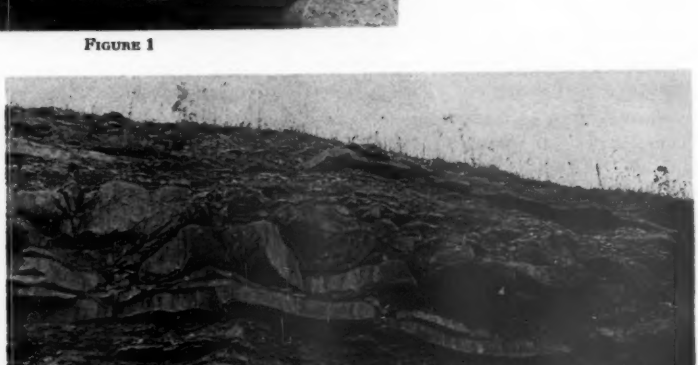

FIGURE 2

FIGURE 3

INTRASTRATAL FLOWAGE FEATURES; INTERSTRATAL FLOWAGE FEATURES; FONDO BEDDING AND AUTO-BRECCIATION IN LIMESTONE

FIGURE 1

FIGURE 2

FIGURE 3

FLUTE CASTS ON UNDERSIDES OF CLINO SILTSTONES AND TWO DISTINCT TYPES OF FONDO BEDDING

FIGURE 1

FIGURE 2

FONDO BEDDING IN SLATE AND IN DOLOMITE

vironment, whereas any shaly material preserved in unda rocks would occur generally as thin laminae on the bedding planes or in small pockets.

Owing to the extensive reworking to which they are subjected, most unda deposits are relatively clean and free from clayey material. Probably almost all the great clean-quartz sheet sandstones (except those obviously eolian) were formed under unda conditions.

On a rapidly sinking undaform where nearby rivers were carrying in large quantities of clay and silt, the accumulating sediments might not have been exposed to reworking by the waves for a time long enough to permit complete winnowing such as is accomplished on the more stable undaforms. Rocks originating under such conditions show bedding features characteristic of the unda environment, but the texture is relatively fine, and the sorting incomplete. Shale is rare or absent. Rocks of this character are probably being formed now in the northwestern part of the Gulf of Mexico. Good examples in older sequences are interstratified in the Hamilton, Portage, and Chemung formations of central and eastern New York.

The fragmental character of unda limestones may not always be evident at first glance, but it is generally clearly visible either on a weathered surface or on a surface polished and then etched lightly with hydrochloric acid.

Any limestone composed largely of broken shell fragments can be ascribed with confidence to either the unda environment or, exceptionally, to the upper part of the clino. No known agencies at work on either the clinoform or the fondoform are capable of fragmenting the shells and then sorting the fragments from the inclosing muds. And except for the upper part of the clinoform, where coarse material is dumped over the outer edge of the undaform, the bits of shell material composing these fragmental limestones could not have been transported in quantity to either the clino or unda environments because such coarse material could not have been carried in suspension.

Though limestone is commonly supposed to be a deep-water deposit, features of texture, composition, and bedding indicate that many limestones are products of the unda environment and therefore were formed in relatively shallow water.

That the unda should be a favorable environment for the formation of limestone follows from the fact that it is the one most favorable for the growth of the larger shell-bearing and

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PLATE 3.—FLUTE CASTS ON UNDERSIDES OF CLINO SILTSTONES AND TWO DISTINCT TYPES OF FONDO BEDDING

FIGURE 1.—FLUTE MARKINGS

On under sides of four different siltstone beds in clino strata at foot of Constitution Hill, Aberystwyth, Wales. Looking up at under sides of beds dipping away from observer. Markings are essentially parallel in all beds. To see pattern of flutings as it was on the tops of the shales before the siltstone was deposited, turn picture bottom-side up. Flutings are believed to have been made by storm-induced density currents of silt-laden water flowing down the slope of a clinoform.

FIGURE. 2.—ORGANIC KEROGEN-BEARING DOLOMITE CONSTITUTING RICH GREEN RIVER "OIL SHALES"

At entrance of U. S. Bureau of Mines experimental mine near Rifle, Colo. Believed to be a fondo-type organic rock. Bedding in fresh rock shown by vague color banding. Note remarkable evenness and massive uniformity of bedding.

FIGURE 3.—FONDO-TYPE CHALK

In quarry near Dunstable, England. Massive bedding, with only one stratification plane recognizable. Megascopic fossils extremely scarce. Scale indicated by hammer.

PLATE 4.—FONDO BEDDING IN SLATE AND IN DOLOMITE

FIGURE 1.—FONDO-TYPE CLAY ROCK, NOW HIGH-GRADE COMMERCIAL SLATE

Penrhyn quarry, near Bangor, Wales. Bedding shown by light streaks and by vague color banding. Scale indicated by small pipe extending down the quarry face.

FIGURE 2.—FONDO-TYPE HUNTINGTON DOLOMITE (SILURIAN)

In quarry about 2½ miles west of Portland, Ind. Main features of bedding are not obscured by dolomitization. Scale indicated by hammer at left.

lime-secreting aquatic organisms. On a stable, but slowly sinking undaform where the quantity of sediment supplied by rivers is not great, the growth of various lime-secreting organisms is seemingly rapid enough to compensate for the sinking and thus to produce great thicknesses of fragmental limestones, all showing the characteristics of the unda environment.

Sedimentary Rocks of Clino Origin

Rocks originating in the clino environment (if we exclude the shelf-edge sands carried by bottom traction from the undaform and deposited at the top of the clinoform) should be expected to show the following characteristics:

Texture—Generally fine enough for the material to have been carried in suspension; mainly that of silt or clay.

Composition—Silt or siltstone and clay or shale or their calcareous equivalents.

Bedding—Regular alternations of silt-sized and clay-sized materials (Pl. 1, fig. 3); bedding relatively thin—siltstones half an inch to 10 inches as a rule, but locally thicker; shales commonly half an inch to several feet; bedding remarkably even and persistent; laminations within the siltstone beds even and regular, without evidence of tractional movement except rarely near the tops of the beds; lack of evidence of bedding features such as are produced by wave action; general, but not universal, absence of cross-bedding and ripple marks.

Flutings and striations—Tops of many of the shale beds fluted and striated by sheet flow of water (density currents) moving down the slope of the clinoform (seen usually as casts on the under sides of the overlying siltstone beds); groovings and gougings roughly paralleling the striations and flutings (probably produced by objects dragged down the slope by density currents).

Intra-and interstratal flowage features—Internal crumplings of the laminations of the siltstone beds, not affecting the tops and bottoms of the beds (Pl. 2, fig. 1); interstratal flowage and crumplings involving more than one bed (Pl. 2, fig. 2).

Among the most striking features on this list are the casts on the bottoms of siltstone beds of flutings and striations (Pl. 3, fig. 1) interpreted as having been produced by sheet flow of density currents. The actual flutings on the shales are seldom seen in outcrop because the shales weather so rapidly, but the casts on the bottoms of the overlying siltstone beds are conspicous among the loose rocks of the stream courses where they occur and are readily found in place on the bottoms of projecting siltstone beds.

These features can almost always be found on rocks that show the type of bedding characteristic of the clino environment.

As soon as the key was found to the interpretation of these markings (Rich, 1950), it became possible to use them not only as criteria for beds deposited on a sloping surface but also for determining the direction of that slope.

During the summer of 1949 the writer tested this interpretation on rocks of the Portage formation near Ithaca, N. Y. Over an area extending about 3 miles east, 5 miles southwest, 10 miles north-northwest, and 18 miles north-northeast, airline distances from that city, he discovered that wherever the bedding interpreted as indicating the clino environment was found, casts of flutes and striations occurred on the under sides of the siltstones. Over the entire area and at stratigraphic horizons including several hundred feet of section, the striations observed nowhere varied more than about 10° to either side of east-west, and wherever direction could be determined definitely, they indicated flow toward the west.[3]

Essentially the same direction was found in the few observations the writer made in clino-type rocks of the Chemung formation of northern Pennsylvania about 45 miles southwest of Ithaca.

These relations the writer interprets as indicating that during Devonian deposition in that region clinoforms were being built westward at several horizons into the sea occupying the Appalachian geosyncline as it was being filled from the east.

At the beginning of this section, exception was made of the coarser sediments worked off the undaform and deposited on the upper part

[3] Sheldon (1929) found practically the same direction of movement indicated by current ripples on the upper parts of the siltstone or fine sandstone beds in the Portage formation of the Ithaca region.

of the clinoform. There, coarser sediments would probably be interbedded with the characteristic alternating siltstones and shales typical of the clino environment. If oversteepening of the top had produced landslips, the exotic blocks, breccias, and conglomerates, and graded bedding described by Kuenen and Migliorini (1950) should be found interbedded with clino and fondo deposits on the lower part of the clinoform and adjacent parts of the fondoform.

The writer has seen the type of intrastratal flowage illustrated in Figure 1 of Plate 2 only in beds that he would interpret as clino deposits, but interstratal flowage of the type illustrated in Figure 2 of Plate 2, in which the individual beds are contorted in places and undisturbed a few feet away, and in which more than one bed may be involved, is common not only in beds interpreted as of clino origin, but also, and in even better development, in beds that are undoubtedly of unda type.

The problem presented by interstratal flowage of the type illustrated in Figure 2 of Plate 2 has, to date, defied solution. The phenomenon seems to be confined to beds of sandstone, siltstone, or limestone that evidently was unconsolidated lime sand at the time of its deformation. It may be a gravity sliding phenomenon restricted to an area close to the edge of the undaform at the time movement occurred, or to the very top of the clinoform. Where such flowage affects beds of definite unda type, it may be an extrusion phenomenon produced in unda beds by unbalanced weight developed as succeeding clino beds at a higher level are built forward over them.

These flowage phenomena definitely occur within the beds and are not produced by the sliding of sandy beds on underlying shales.

Sedimentary Rocks of Fondo Origin

Complete quiet or only faint movement of the water and fineness of the sediment supplied are the most outstanding characteristics of the fondo environment. Another is remoteness from the sources of coarse terrigenous sediment. Such sediments ordinarily must settle before reaching the fondo environment. These considerations dictate the following criteria:

Texture—Universally fine, except for the remains of bottom-dwelling shell-secreting organisms and of pelagic forms that live in the overlying waters; fossil remains generally scattered through the rock without having been sorted by water movement (Pl. 2, fig. 3) though, as lag concentrates, fossil shells may have been concentrated locally as a result of contemporaneous non-deposition of the finer sediments in which they would normally have been bedded; limestones mainly either fine-textured oozes (*e.g.* globigerina ooze) or "dense", "lithographic limestone" types (Pl. 2, fig. 3).

Composition—Generally clay, shale, or calcareous or siliceous oozes composed of impalpably fine detrital lime flour, of precipitated lime, or of foraminiferal, diatomaceous, or radiolarian accumulations, all mixed with clay in various proportions; locally interbedded clay and silt or their lithified equivalents; less commonly, bituminous muds or shales or calcareous organic rocks such as some of the Green River "oil shales" which consist essentially of dolomite and kerogen (Pl. 3, fig. 2).

Bedding—Even, generally tending toward massive (Pl. 3, fig. 3; Pl. 4, figs. 1, 2); may be even and thin-bedded (Pl. 3, fig. 2); locally (probably near the base of a clinoform) interbedding of clays and shales with thin, evenly bedded layers of silt without flutings or striations; bedding commonly inconspicuous, but generally revealed in a large exposure by vague color banding (Pl. 3, fig. 2; Pl. 4, fig. 1) and, even in small exposures, by weathering.

Organic rocks such as the oil shales are here interpreted as products of fondo deposition in unaerated basins where the water becomes toxic enough to kill bottom scavengers and thus to permit the preservation of organic material.

Shale is believed to be one of the most common and widespread rocks formed in the fondo environment. The very fact that a marine, lagunal, or lacustrine rock is a shale (provided it is not interbedded with siltstones showing flute casts, as in the clino environment) is believed to be good presumptive evidence of fondo origin, because material of texture fine enough to produce shale would not ordinarily remain on the undaform and normally cannot come to rest permanently until either the clino or the fondo environment has been reached. If such material had settled on the clinoform it would

be interbedded with siltstones showing flute casts and striations. Consequently, any thick, homogeneous mass of shale is strongly suggestive of a fondo origin.

Interbedded shales and thin siltstones should be expected in that part of the fondo environment nearest the base of the clinoform because, during storms, the muddy water of the density current flowing down the clinoform would normally spread widely before its silts could settle. Siltstones so formed would not carry flute casts or striations, however, because the velocity of the density current would have been checked when it reached the base of the clinoform, and its capacity for scouring the underlying clays would have been lost. The siltstone layers would become progressively thinner and lie closer together with distance away from the base of the clinoform. In this way, alternating beds of silt and clay of almost paper thinness might be produced in the regions more remote from the clinoform.

A common feature shown by many limestones of the dense, "lithographic" type and by very rich bituminous shales, is shrinkage cracking exhibited both on the surface and in cross sections. Such polygonal shrinkage cracks in the dense limestones, and also in the Green River oil shales, have commonly been interpreted as sun cracks, but the writer suggests that they were produced by subaqueous or subterranean shrinkage associated with the expulsion of water during compaction of a substance which tended to set and become brittle before shrinkage due to expulsion of water was complete. (*Compare* Richter, 1941; Twenhofel, 1923.)

These "dense" types of limestone not uncommonly show a peculiar brecciation (Pl. 2, fig. 3) which seems to have been produced in place, possibly by shrinkage after the surface of the ooze had set hard enough to break, but before compaction and consolidation of the rock had gone far enough to prevent limited flowage movements. Brecciation ranges from mere fragmentation with only slight relative displacement of the fragments to complete disruption, with the formation of limestone breccias indicating contemporaneous movement of small or larger scale.

In general, flowage phenomena are not expected in the fondo environment, but, if the surface on which the sediments are being deposited has considerable topographic irregularities, or if tectonic activity produces them during deposition, flowage might readily occur in material as fine and unstable as water-soaked lime oozes.

Though dolomites make up a considerable proportion of the calcareous rocks in the geologic column, they are not specifically considered in this discussion because dolomitization is commonly secondary and may affect limestones formed in any of these environments. Furthermore, during secondary dolomitization, the broader features and bedding of the rock are not ordinarily obscured and would persist as clues to the environment in which they were deposited (Pl. 4, fig. 2).

General Considerations and Applications

Deceptive Thickness of Clino Sequences

The apparent thickness of a series of rocks constituting a clinothem, if measured in the ordinary way by taking thicknesses at right angles to the bedding at successive exposures, would always be greater than the true thickness of the unit, and the error would increase with increase in the angle of slope of the clinoform at the time the beds were deposited—*i.e.*, with the original dip of the clino beds. The true thickness should be measured at right angles to the base or top of that unit, not at right angles to the bedding of the individual clino beds. (*See* Figure 2-A.)

This possibility for error was pointed out long ago by T. C. Chamberlin (1914b) and by Cotton (1918). It is self-evident and can readily be avoided if one recognizes the beds which he is measuring as being of clino origin.

Undaform-Edge Sands

Under stable conditions, wave erosion extends the undaform landward while deposition at its edge extends it toward deep water. The base of the undathem—*i.e.*, of the body of coarse sediments deposited on it at its outer edge—would ordinarily be relatively flat, but jagged in detail as a result of pendent sheets of the coarse unda material projecting for varying distances down onto the upper parts of the

clinoform as it was being built forward. (*See* Figure 2-B.)

The extent of such downward-projecting irregularities of the base of the undathem would seem to depend mainly on the regularity in the ordinarily be coextensive with that of the edge of the undaform, and its thickness measured at right angles to the dip of the clino beds would depend on factors such as the thickness of the coarse material on the undaform before the

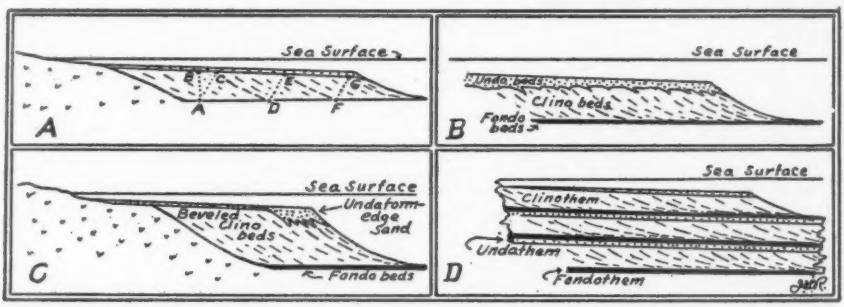

FIGURE 2.—SKETCHES ILLUSTRATING CONCEPTS DISCUSSED IN THE TEXT

(A) Sketch showing why the apparent thickness of a clinothem as measured from outcrop sections may be much greater than the true thickness. True thickness is A-B; thickness as measured might be A-C plus D-E plus F-G.

(B) Sketch cross section showing downward interfingering of coarse unda beds with finer clino beds to produce an irregular base of the undathem.

(C) Sketch showing zone of thickened sand at outer edge of undaform (undaform-edge sand) produced by slight lowering of sea level.

(D) Sketch showing alternations of fondothems, clinothems, and undathems such as would be produced by intermittent rise in sea level or sinking of the land.

Vertical scale greatly exaggerated in all sketches.

forward building of the undaform. A steady rate of advance of its edge would tend toward regularity. Under relatively uniform conditions, this process should not ordinarily produce any great variations in the thickness of the undathem.

If, however, after an undaform had been well developed and graded to a particular level of the water body, that level were lowered, the waves would start to grade the surface of the undaform to the new and lower level. In doing so, they would transport great quantities of the coarse, unconsolidated material lying on the floor of the undaform out to its edge comparatively rapidly and would dump it over the edge to form a belt of thickened sand-textured unda material at the top of the clinoform in a zone extending all along the edge of the undaform (Fig. 2-C).

The writer proposes to call such a belt of sand-sized material an ***undaform-edge sand***. It might be composed of quartz sand, lime sand, or a mixture. It might be 100 or more feet thick, measured vertically; its length would change in sea level occurred, and on the extent to which this was reworked in the process of grading it to the new profile of equilibrium. The width of the undaform-edge sand belt would depend on the above factors, on the depth of water off the edge of the undaform, and on the original angle of slope of the clinoform. It might be very considerable.

Undaform-edge sand bodies are in excellent position to be preserved by burial under later sediments. All that is necessary for their preservation is a halt in the lowering of the water level and its rise again later or, even, merely a regional sinking of the area.

Undaform-edge sand bodies are especially well situated to become reservoirs for the accumulation of oil and gas, provided later structural or erosional and depositional events contribute to that end. They lie at the upper limit of clino deposits consisting of alternating muds and siltstones or fine sands. Of these, the muds are likely to contain abundant organic source material for oil, and the silts, in their role as "carrier beds" for upward-moving compaction

water, could readily carry the oil and gas upward toward the reservoir.

The writer suggests that some of the supposed shoreline trends along the Gulf coast and elsewhere may have had such an origin. Certainly undaform-edge sand bodies are in much more favorable position to be preserved than are sand bars formed at the seashore.

A suggested criterion for the recognition of undaform-edge sand bodies is that they should finger downward from the main sand body between alternating shales and silts of the clino type, and their tops should be essentially flat. If the strata have subsequently been tectonically disturbed, the original attitude of the top of the sand body might be determined by comparison with the bedding of adjacent definitely recognizable unda strata, which could confidently be assumed to have been deposited at very low angles.

The original downward fingering of an undaform-edge sand body into the silts and shales of the clinoform might appear, after deformation had obscured its dip, to be merely lateral gradation of the sands into silts and shales. But that, in itself, would be a good criterion for an undaform-edge sand body, or at least for its deep-water edge, because an off-shore sand body or a strand-line sand body should grade seaward into and interfinger with unda-type deposits rather than into the alternating silt and shale beds of the clino type.

Effects of Changes in Relative Level of Land and Sea: The Building of a Continental Shelf

Changes in water level relative to the land would cause great variations in the proportions of the total sediment load deposited on the undaform and on the clinoform, respectively, but relatively little change in that made on the fondoform.

The effects of various changes in level of land and water can readily be deduced by the reader, and need not be detailed here. In the course of the building of a continental shelf, fluctuations of sea level due to causes other than local diastrophic movements are commonly superimposed on a general sinking, or a sinking and seaward tilting of the entire coastal region.

If we start with a broad continental shelf whose floor is a graded undaform extending all the way out to its edge, a rise in sea level will cause a rise of wave base and will start the grading of a new undaform at a higher level. Meanwhile, on the outer part of the shelf, which will now be below wave base, sediments of fondo type will be deposited over those of the former undaform surface. (*See* Figure 2-D.) As the new undaform is built seaward, clino beds, and above them unda beds, will be laid down over the new fondo sediments on the outer part of the shelf.

If these processes were to be repeated many times and were complicated somewhat by beveling of former deposits whenever sea level temporarily dropped especially low and if, meanwhile, the entire region were sinking slowly so that most of the products of these fluctuations would be preserved, a great continental shelf could develop, thousands of feet thick and composed of individual undathems, clinothems, and fondothems of various thicknesses superimposed on each other (Fig. 2-D). The individual tripartite sequences need not be thick, for minor fluctuations in sea level would be sufficient to produce them. Completeness of any single sequence would depend on the vicissitudes of sea level change.

The final continental shelf so built up would be a composite product and not a single delta-like unit built steadily forward into the sea.

During the building of such a composite shelf, any considerable lowering in the level of wave base would cause removal of material previously deposited on the undaform and, possibly, beveling of the underlying clino or even fondo deposits. Unconformities of the type so produced must be very common. If clino beds had been beveled, the unconformity would show an angular discordance equal to the angle between the essentially flat base of the undathem and the depositional dip of the clino beds (from a small fraction of a degree to at least 6°), but such discordance would not connote emergence of the land nor crustal movement of a type that would cause changes in dip.

Facies Considerations

The nature of the sediment deposited at any one place depends less on the total make-up

of the sediment being supplied from the land than it does on the environment in which it is being deposited. Any idea that shale or limestone, for example, indicates low land in the source area and sandstone indicates relatively high land should be most carefully scrutinized.

Conditions for life on the undaform, clinoform, and fondoform are very different, yet, in a given body of water, all these environments exist at the same time adjacent to each other. On the undaform, the relatively clear, well-aerated water, with light penetrating nearly or quite to the bottom, is favorable to abundant life of shellfish or reef types. On the adjacent clinoform the oft-repeated showers of mud and its relatively rapid accumulation must be unfavorable for such life, and only forms especially adapted to those muddy conditions, or the remains of forms that lived in the surface waters would be found. On the relatively deep fondoforms mud would not be settling to excess, and forms adapted to life in the depths where no light is present would prevail, along with the remains of pelagic forms that had settled from the waters above. In un-aerated depressions, the toxic-water organic shale facies would also be common. In a fondo environment shallow enough for light to penetrate, life undoubtedly would be abundant wherever the water was not toxic.

Thus three different but contemporaneous facies faunas would exist, and, if not recognized for what they are, they might cause endless confusion in correlations.

Since the evolution of life is likely to be slow in comparison with the rate of sedimentation on the undaform and clinoform, there should ordinarily be greater differences at a given time between the faunas of the unda, clino, and fondo environments than there would be in any one of them over a long period of time.

Paleogeographic Applications

Ability to recognize rocks deposited in the unda, clino, or fondo environments by their physical and faunal characteristics should make it possible to reconstruct the geologic history and paleogeography of a region in much greater detail and with more exactness and greater confidence than heretofore. This is particularly true where flute markings on the clino beds make it possible to determine in which direction the clinoform was sloping at the time of deposition.

A few examples of such determinations made recently in reconnaissance illustrate the point:

In Cascadilla gorge at Ithaca, N. Y., a section from the College Avenue bridge downstream to the mouth of the gorge reveals excellent examples of undathem, clinothem, and fondothem. The upper 150 feet, approximately, of that section is an undathem composed almost entirely of fine sandstone and coarse siltstone, with thin wavy bedding of the type shown in Figure 2 of Plate 1, though not so knobby. Scarcely any shale is present, and no flute markings were found.

Below, starting beneath the Stewart Avenue bridge, is a clinothem 50 to 75 feet thick with the typical alternation of siltstone beds roughly 2 to 6 inches thick with considerably thicker beds of clay shale. Most of the siltstone beds carry well-developed flute markings and striations on their under sides. The direction of these striations was estimated to be very close to east-west, with flow from the east.

Next in order down the gorge is a fondothem (the Renwick shale) composed of clay shale in which no siltstone beds were found.

Thus the exposures in this gorge represent a complete cyclic unit consisting of a fondothem at the base, followed upward by clinothem and undathem. The geological history indicated by this sequence would be: first, a relatively deep sea with its bottom below the reach of wave action, in which clay was being deposited. In time a clinoform was built forward over it from the east as indicated by the flute-marked siltstones interbedded with clays. If no subsequent beveling of the clinothem occurred, we can deduce that the clinoform was built forward into water whose depth was 50 to 75 feet below wave base, for that is the true thickness of the clinothem. After filling to wave base, unda conditions, with full wave working of the bottom (probably contemporaneous with slow sinking of the region), lasted long enough for the deposition of at least 100 feet of unda strata.

Reconnaissance elsewhere higher in the section shows other similar and, in some instances thinner, undathems, clinothems, and fondothems, all within the Portage formation.

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Thus in this region the Appalachian geosyncline was not filled by the advance of a single delta-like unit but, rather, by the successive advances of many unda and clino units recording many fluctuations of sea level superimposed, probably, on a general sinking of the region that permitted the various units to be preserved.

Starting at the bottom and working up the column in the order in which the strata were deposited, use of the criteria developed in this paper leads to the following tentative interpretations for the Ithaca region:

In the stream gorge entering Cayuga Lake near the cement mill at Myers (south of Ludlowville), the lower 40 to 50 feet of the Hamilton formation is a typical fondothem. Above it about 2 feet of crinoidal limestone, fragmental and thoroughly sorted, must be either an undathem or, possibly, a lag concentrate. Its clean-cut base suggests the former, and also that it is a product of temporary lowering of wave base, possibly attended by removal of considerable previously deposited material. The next 10 feet is transitional between undathem and fondothem with several thin beds of shell-hash limestone alternating with shales. Then comes more than 75 feet of fondo shales followed by the Tully limestone which also is a fondothem. The Tully grades upward into the toxic-water Geneseo bituminous shale of fondo type. The overlying Sherburne flags were not examined. Above them comes the section already described for the Cascadilla gorge. Higher in the Portage come alternating fondothems, clinothems, and undathems, with the latter appearing to compose by far the larger part of the section. The proportion of undathem in the section appears to increase to the east.

From the above description, the reader can readily interpret the paleogeographic history.

Another area further illustrates the use of our criteria in paleogeographic interpretations—the Ordovician section in the gorge of Kentucky River at the crossing of Route 25 south of Lexington. (For the section, *see* McFarlan, 1943.)

Beginning at the bottom of the section, the Camp Nelson, Oregon, and Tyrone members of the High Bridge limestone are all undoubted fondothems. The overlying Curdsville, Hermitage, Jessamine, and Benson members of the Lexington limestone are all clearly undathems, composed of shell-hash and showing characteristic bedding like that of Figure 2 of Plate 1. The overlying Brannon, however, shows features of the clino type. It consists of strikingly even bedded alternating mudstones and limey siltstones. The siltstones occur in beds up to 6 inches thick separated by 18 inches to 2 feet of shale. No flute markings were seen, but many of the thicker silty beds show prominent intrastratal flowage phenomena. These rocks are interpreted as having been formed low on the clinoform or on the fondoform near its base where the inclination was too low to impart erosive power to density currents. The top member of the Lexington—the Woodburn—is an unmistakable undathem with practically no shaly material.

The lower half of the Cynthiana formation overlying the Woodburn is transitional; its bedding is of clino type, while the composition of its non-shaly beds is partly silt, suggesting clino environment, and partly shell hash, suggesting unda environment. Groups of beds containing the siltstones alternate with those containing shell hash in units roughly 10 feet thick. Many of the non-shaly beds show flowage features, and some of them have been pulled apart during the flowage, producing gaps a few feet southeast of the crumpled sections. This member is interpreted as formed under transitional conditions in the extreme upper part of a clinoform whose slopes were very gentle, and it seems that periods of true clino conditions alternated with others in which the deposition of mud was intermittently interrupted by the waves of exceptionally severe storms reworking the bottom.

The remainder of the section as there exposed, including the upper half of the Cynthiana formation, the Million shale member of the Eden formation, and the Garrard siltstone member of the Maysville formation, shows oft-repeated alternations of fondo and unda features such as have been described for the Ordovician rocks of the Cincinnati region. The Garrard siltstone at the top of the section has rather massive bedding and is strongly flow-rolled. It is tentatively interpreted as a fine-grained undaform-edge sand.

These interpretations show that the criteria can be used with the greatest confidence for many rock units, but that transitional units

may have some of the characteristics of two environments. Even in the transitional units, however, much can be determined about the conditions of deposition.

Conclusions

This paper records an attempt to develop criteria by which rocks formed in the unda, clino, and fondo environments can be recognized and differentiated.

Rocks formed in the unda environment—that lying above wave base—show features produced by wave and current agitation peculiar to that environment. These are: distinctive wavy bedding, cross-bedding, ripple marks, coarse texture, and sparsity of clay-sized material.

Rocks formed in the clino environment—that comprising the slope from wave base down to the floor of the water body—show regular, thin interstratifications of silt- and clay-sized material, and aligned flow-markings and flute casts and other flowage phenomena indicating deposition on an inclined surface. They are generally, but not always, free from ripple marks and other indications of agitation of the water, other than that of subaqueous sheet flow believed to be caused by density currents.

Rocks formed in the fondo environment—that representing the generally flat floor of the water body—are fine-grained, with even, generally massive bedding and are generally free from flow markings, ripples, or other evidences of agitation of the water, though exceptionally, ripple marks may be present. Clays and oozes are characteristic.

Reconnaissance application of these criteria proves their pertinence and value.

Establishment and perfecting of such criteria will make it possible to read in the rocks the history of the paleogeographic environment at the time of their deposition much more fully and accurately than has heretofore been possible.

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HIGHLAND CENTERS OF FORMER GLACIAL OUTFLOW IN NORTHEASTERN NORTH AMERICA

By Richard Foster Flint

Abstract

Review of the available evidence indicates that several highlands in northeastern North America stood high enough to maintain local glaciers after they emerged from beneath the general ice sheet. Local glaciers probably preceded the ice sheet on most or all of these highlands. The implied glacial history is more complex than has been generally realized.

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Introduction

The glaciation of North America east of the Rocky Mountains was long thought of in terms of one or two very extensive ice sheets having large and rather vague central areas of dispersal and, in the peripheral areas, rather rigid unvarying directions of flow. The corresponding deglaciation was visualized as the simple migration of a glacier terminus inward toward the vague central region.

The gradual recognition of cirques in some of the highest mountains in northeastern United States and eastern Canada forced local modifications of this concept. In general, however, the concept of local highland ice caps not related to cirques has received very little attention, despite evidence of such ice bodies early adduced by Ells, Chalmers, L. W. Bailey, and others, and despite the published opinion of Tarr that ice bodies of this kind probably existed.

It has long been well established that in Britain, during the latest glacial age, there were several highland centers of radial outflow of ice, and that with fluctuation in the climate these centers waxed, waned, merged, or became detached from each other in a complicated way. Similarly the ice sheet in Scandinavia has been shown to have had several centers of radial outflow that shifted position with the passage of time. (Cf. Frödin, 1925; Ljungner, 1949.)

The writer, believing that the prevailing concept of the former ice was rigid and improbable, critically studied the literature and concluded that a more flexible hypothesis, involving several centers of radial outflow of ice, shifting in position with time, was justified

(Flint, 1943). Field examination of a critical district, with this problem specifically in view, led to the conclusion that in the highlands of Gaspe the general ice sheet was succeeded first by a local center of radial outflow of ice, and later by a set of valley glaciers originating in cirques (Flint, Demorest, and Washburn, 1942).

Believing that a similar history may well apply to other highland areas in eastern North America, the writer has attempted to assemble the facts and opinions on the general subject of local centers of outflow within the general ice sheet, and of local glaciers in highland districts both before and after the ice sheet covered them. In some cases definite conclusions can be drawn as to the sequence of glacial events. In others the evidence is inconclusive because of insufficiently detailed field study. The results, however, seem to justify retention of the concept of a number of local centers, pending the accumulation of further evidence. It is hoped that the present summary will be useful as a guide to the literature and as a stimulus to further observation.

Each of the highlands, known or believed to have harbored local ice during the latest glacial age, is considered. The list of them is of course very incomplete, because the glacial geology of the far northeastern part of the continent (including the Arctic islands) is almost unknown. Future exploration will undoubtedly reveal additional highlands from which ice flowed outward in the past, besides those that possess local glaciers today.

In the central and western parts of the continent, but still east of the Rocky Mountains, there is scattered evidence of other centers of outflow of ice. These centers, if centers they are, differ from those farther east in that they occupy lowlands and must therefore have been localized primarily by the topography of the glacier ice rather than by the topography of the ground beneath the ice. They are not discussed in this paper.

1[1]. Catskill Mountains

The Catskill Mountains (actually a much-dissected plateau) lie in eastern New York State, centering near 42°00′ N. Lat. and 74°30′ W. Long. Their highest parts slightly exceed 4000 feet in altitude. Abundant evidence indicates that this district was covered by the Laurentide Ice Sheet, though not necessarily during the later part of the Wisconsin glacial age. Glacial striations occur on the summit of Slide Mountain (4205 feet), the highest point in the Catskills (Antevs, 1932, p. 9).

Although most of the valley heads in this high mass are the result of normal erosional processes, a few are ideally theater-shaped and have, upon their floors, small crescentic end moraines. These are unquestionably cirques (Rich, 1906; Johnson, 1917, p. 549–551). Some of them are shown, very poorly, on the Slide Mountain Quadrangle of the U. S. Geological Survey. The features are small, and the undisturbed moraines of local origin show that the occupation of the cirques by local glaciers postdates, at least in part, the uncovering of the district from beneath the ice sheet.

The low altitudes of the cirque floors—somewhat less than 2000 feet—are noteworthy. Possibly they reflect the low temperatures expectable immediately outside the margin of the ice sheet early in the deglaciation process, while the ice was still thick.

2. Adirondack Mountains

The Adirondack Mountains are an irregular, monadnocklike group in northeastern New York State. About 10 high summits reach altitudes close to 5000 feet. These center near 45°00′ N. Lat. and 74°00′ W. Long. On Mount Marcy (alt. 5344 feet) unquestionable foreign stones of glacial origin occur up to altitudes of slightly more than 4900 feet (Robert Balk, personal communication). Thus there is direct evidence that the Adirondacks were nearly buried beneath the Laurentide Ice Sheet. That they were completely buried is suggested by the fact that the anorthosite from which Mount Marcy is sculptured is generally fresh and undecomposed right up to the summit of the mountain. As anorthosite decomposes readily under the climatic conditions prevailing in the region, the lack of decomposition suggests relatively recent glaciation of the entire mass.

[1] Numbers refer to locations on the map, Fig. 1.

Several of the higher peaks have cirques incised into them (Ogilvie, 1902, p. 406; Johnson, 1917, p. 547–549). The cirques are poorly developed (on the Lake Placid Quadrangle they are not indentifiable as such although they are clearly visible from an airplane flying at 6000 feet), suggesting that the glaciers that excavated them were small and short lived. Undisturbed crescentic end moraines, like those in the Catskills, indicate that the local glaciers postdated, at least in part, the ice sheet.

Altitudes of the cirque floors are mostly in excess of 4500 feet. This altitude, more than twice that of the Catskill cirques, implies[2] that, by the time the shrinking ice sheet had uncovered the Adirondacks, the regional temperature had risen considerably above that which had earlier characterized the emerging Catskills.

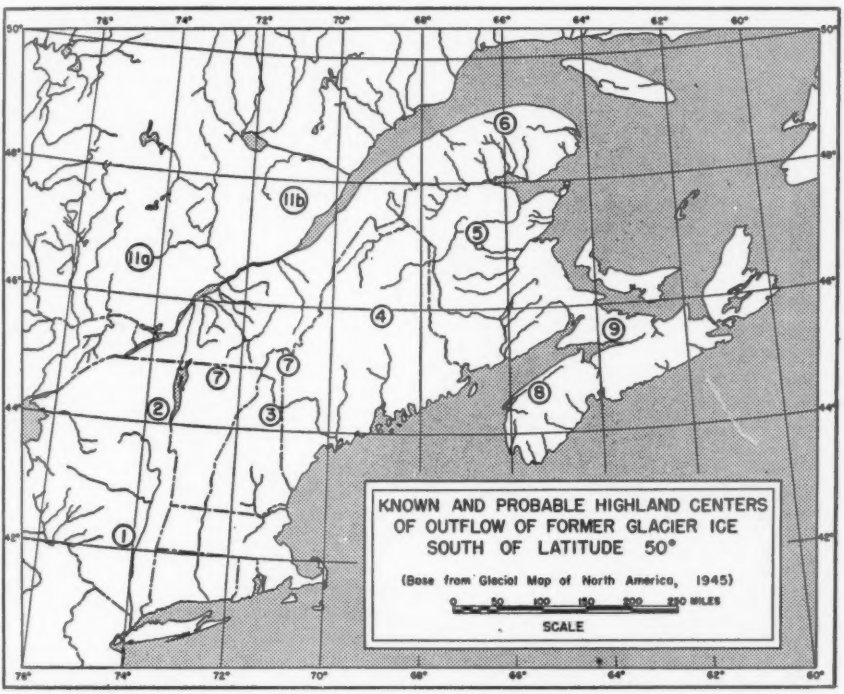


FIGURE 1.—KNOWN AND PROBABLE HIGHLAND CENTERS OF OUTFLOW OF FORMER GLACIER ICE SOUTH OF LATITUDE 50°.

1.—Catskill Mountains
2.—Adirondack Mountains
3.—Presidential Range, White Mountains
4.—Mount Katahdin
5.—Central highlands of New Brunswick
6.—Shickshock Mountains
7.—"Appalachian" center (2 areas shown)
8.—Upland of southwestern Nova Scotia
9.—Cobequid Mountains
11a, 11b.—Local centers within the highland of Quebec-western Labrador
Note: Nos. 10, 11, and 12 are shown on Figure 2.

3. PRESIDENTIAL RANGE, WHITE MOUNTAINS

The White Mountains culminate in the Presidential Range, a great granitic mass standing

[2] The discrepancy may be the result, in part, of differential crustal uplift caused by removal of the weight of the ice sheet since the cirques were made. However, this factor is probably not large.

high above its surroundings. Mt. Washington, the highest peak, with an altitude of 6288 feet, lies near 44°20′ N. Lat. and 71°20′ W. Long.

Till, erratic stones, and striations indicate that even this highest summit was buried beneath the Laurentide Ice Sheet, apparently during the Wisconsin Glacial age.

In addition, the heads of at least 10 valleys on the Presidential Range are modified by cirques (J. W. Goldthwait, 1913; R. P. Goldthwait, 1940, p. 12–25), with floor altitudes in the neighborhood of 4000 to 4500 feet. Undoubtedly both the latest cirque glaciers and the latest ice sheet in this district date from the Wisconsin Glacial age. However, two mutually opposed views have been held concerning the relative dates of cirque glaciers and ice sheet within that age. J. W. Goldthwait (1913) and R. P. Goldthwait (1940) adduced evidence that the cirque glaciers were the earlier. Johnson (1917) considered them the later, but modified this view (Johnson, 1933) after Antevs (1932) had shown that, although glaciers occupied the cirques after they were uncovered from beneath the ice sheet, the cirques were probably cut chiefly before the advent of the Wisconsin ice sheet. The evidence consists of a large discrepancy between the volumes of the cirques and the volumes of drift immediately below them. Antevs thought it likely that cirque excavation had commenced early in the Pleistocene epoch and had been continued during each glacial age. The writer concurs.

Thus during the Wisconsin Glacial age the Laurentide Ice Sheet probably was both preceded and followed by cirque glaciers in the White Mountains. This concept was foreshadowed in a suggestion by Tarr (1900, p. 446–447), who implied an early transition from cirque glaciers through a local ice cap into the ice sheet itself, with the reverse evolution occurring during deglaciation. Upham (1904) cited specific evidence (in the trends of end moraines) of such an intermediate ice cap. However, J. W. Goldthwait (1916a, p. 265) rejected this evidence on the ground that it did not include proof of the northward transport of any drift.

In summary, it can be concluded only that in the White Mountains the latest ice sheet was both preceded and followed by cirque glaciers. Reliable evidence is lacking that the transition from cirque glaciers to ice sheet, and vice versa, took place through an intermediate local ice cap. However, there is clear evidence of such an intermediate ice cap in the Shickshock Mountains, and some evidence that ice caps existed (without the cirque glaciers) in Nova Scotia. Meteorologically Mt. Washington is the best place in New England for the accumulation of snowfall. The weather at the summit is more severe than on any other summit in eastern North America and is comparable with the weather reported from the Antarctic Continent (Brooks, 1940).

4. Mt. Katahdin

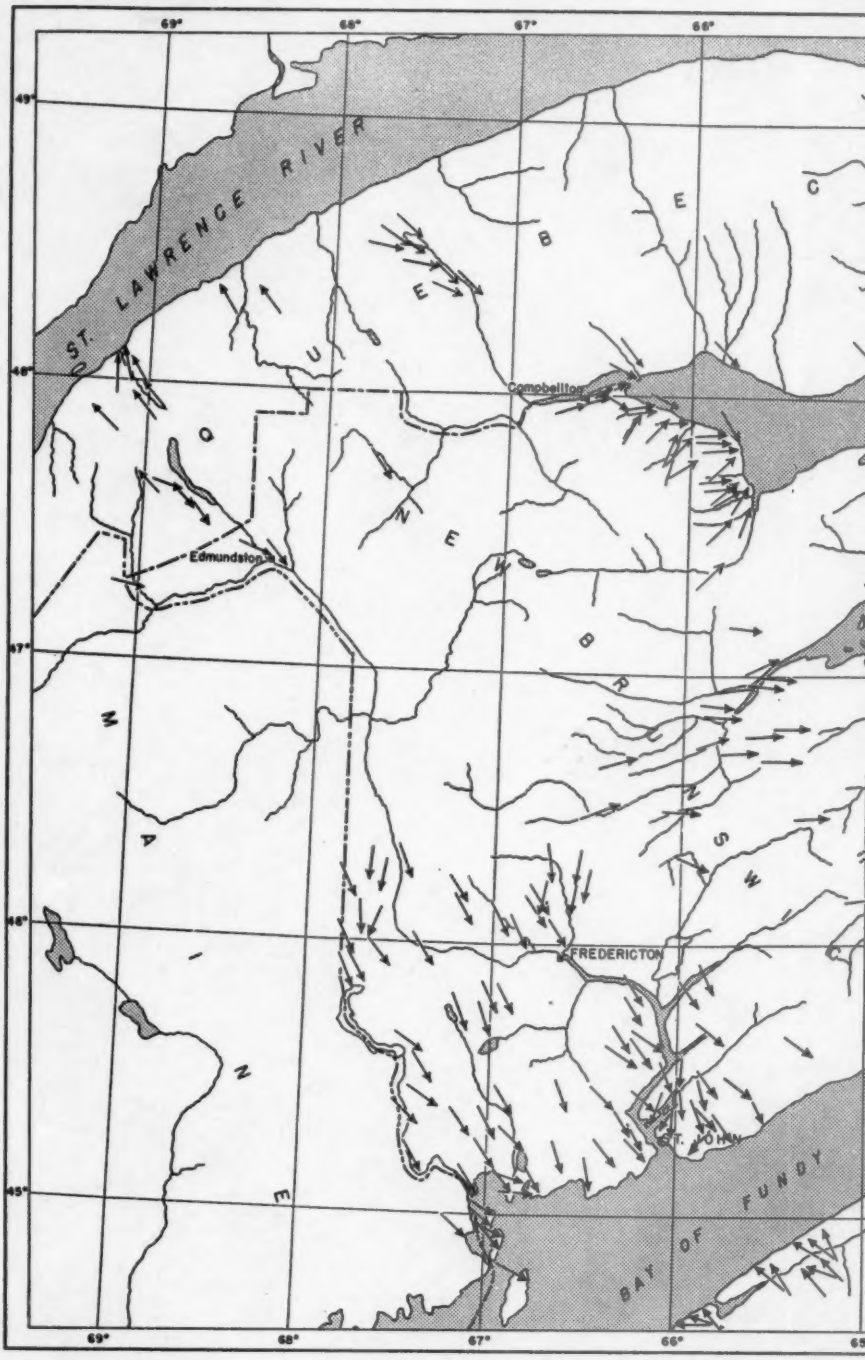
Mt. Katahdin, the dominating peak of a massive monadnock group having a plateaulike top, centers near 46°00′ N. Lat. and 69°00′ W. Long. in north-central Maine. Its altitude is 5267 feet. The area is well shown on the Katahdin Quadrangle of the U. S. Geological Survey. Foreign stones of glacial origin occur on the slopes and summit of the mountain, proving that this mass was overtopped by the Laurentide Ice Sheet.

Large, well-developed cirques are present with floors at 3000 to 3500 feet. Tarr (1900), though he did not specifically identify any forms as cirques, concluded from the forms of the valleys and from extensive local moraines that significant local glaciers had flowed outward from Mt. Katahdin, after the uncovering of this mountain mass from beneath the ice sheet. He implied that, at some time after the general deglaciation, ice had flowed radially down from the plateau, at around 5000 feet, into the cirquelike basins at a much lower altitude. The implication is that a mountain ice cap succeeded the ice sheet and preceded the cirque glaciers.

Antevs (1932, p. 16–22) specifically identified several cirques and reinforced Tarr's opinion that cirque glaciers postdated the deglaciation of Mt. Katahdin by the Laurentide Ice Sheet late in the Wisconsin Glacial age.

5. Central Highlands of New Brunswick

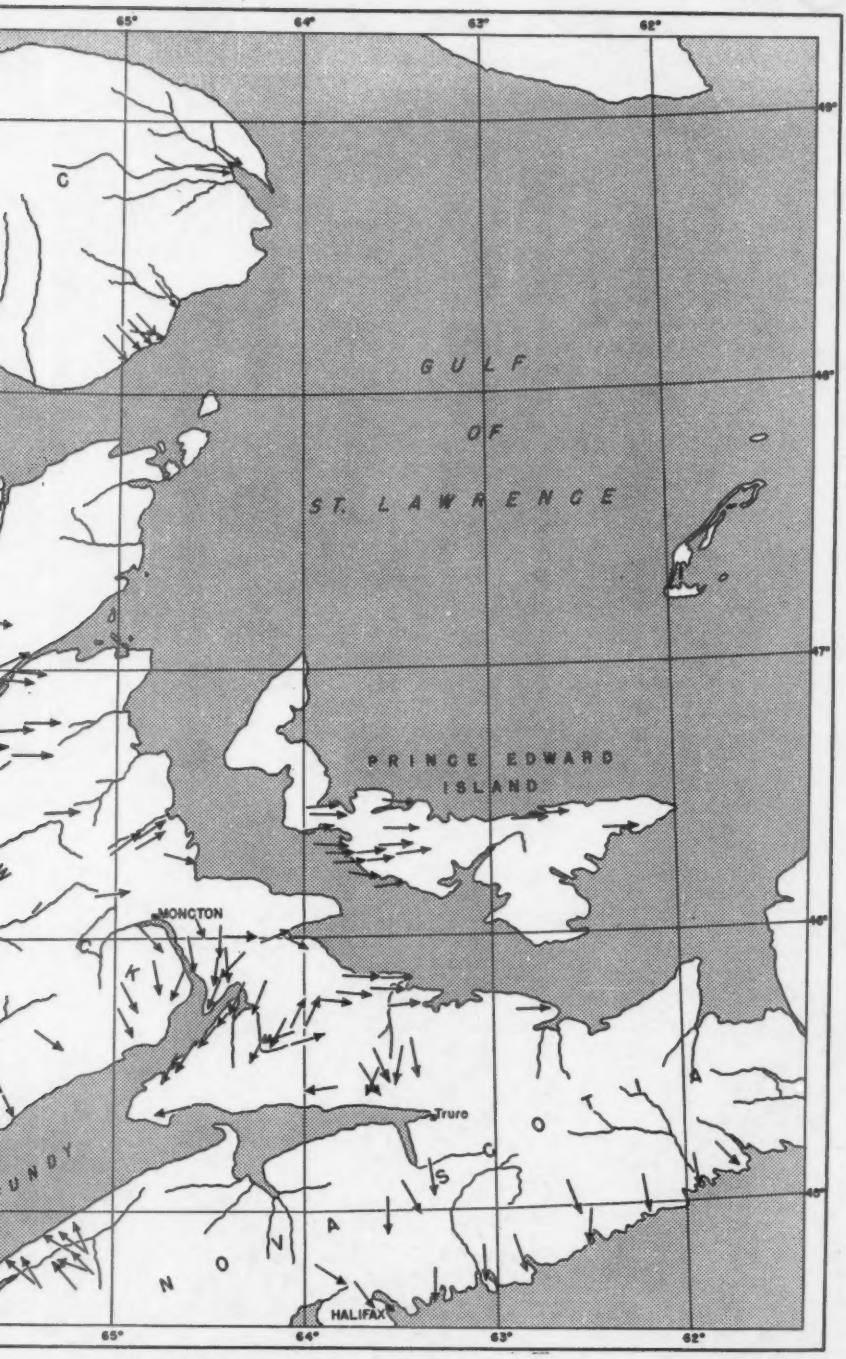
The northern interior region of New Brunswick includes a high plateaulike mass commonly known as the Central Highlands, centering near



STRIATIONS IN NEW BRUNSWICK AND ADJ

—▶ Striations recorded by Chalmers (1

—→ Striations recorded by Alcock



K AND ADJACENT CANADIAN REGIONS

by Chalmers (1895, maps 558, 559).

led by Alcock (1935, p. 121).

47°00′ N. Lat. and 67°00′ W. Long. The highest summit in this group is Mt. Carleton, with an altitude of 2689 feet. The area is not represented on large-scale topographic maps.

No cirques have been reported in these highlands, and none have been inferred. The local glacier ice is thought to have consisted of a highland ice cap that covered the summits and that did not therefore excavate cirques. The former presence of a glacier of this kind is inferred from many striations and from related stoss-and-lee topography. These features (Pl. 1) radiate northward and eastward from the central highlands and are particularly evident along the coast. They were first reported by Chalmers (1886), who inferred a local center of glacial flow. Chalmers' striations, which are represented both in maps and in tables, are impressive, in that in most cases he has distinguished whether or not the specific direction of flow is indicated.[3] South of the highlands, the striations bear southward, but these are not critical because this direction of flow would have been assumed by glacier ice regardless of whether it originated in New Brunswick or were part of the great Laurentide Ice Sheet.

After extensive further study in the region Chalmers (1895) recorded many additional striations and stated that he regarded the hypothesis of a local New Brunswick glacier as established (p. 88M). However, he apparently regarded the local glacier as an alternative to general glaciation, rather than as an addition to it. Undoubtedly this was because the scanty evidence of general glaciation of the high Gaspe Peninsula, to the north, had led him to visualize the Laurentide Ice Sheet as having failed to surmount that high country—as having, in fact, extended no farther southeast than the St. Lawrence River.

After several years' field study of Gaspe and New Brunswick Alcock (1935, p. 125) accepted Chalmers' local New Brunswick ice cap and cited evidence of general glaciation likewise. He considered that the local ice had antedated the Laurentide Ice Sheet, which later merged with and overwhelmed it. He did not, however, detail the evidence from which this relative dating was inferred. Striations recorded by him, in addition to those plotted by Chalmers, are shown in Plate 1.

Alcock (1948, p. 12) found evidence in the unsymmetrical shapes of glacially eroded bedrock bosses that ice had flowed northeastward in northeastern New Brunswick and southward in southern New Brunswick. He concluded that local glaciation originating in the Central Highlands had been important and thought that such local ice existed both before and after New Brunswick was covered by the general ice sheet, the earlier local glaciers having accomplished more work than did the later ones. He noted further that in the Central Highlands glaciation was very light; in places there is no evidence of glaciation at all. This he explained as the result of expectable slight movement at the center of the local ice cap. He suggested that the local ice, when later overridden by the Laurentide Ice Sheet, acted as a mat that protected the underlying rock surface against erosion.

As a result of detailed geologic mapping of northwestern Prince Edward Island Owen (1949) found erratic boulders derived from the mainland of New Brunswick and concluded that the latest glacial movement in that district was from west to east.

The writer believes the striations recording local ice were made after the predominant influence of the Laurentide Ice Sheet had ceased. This opinion is based on the abundance of these striations, which are not likely to have survived a general glaciation characterized by flow in quite other directions. On the other hand, the altitude of the Central Highlands of New Brunswick, and the cold climate that probably enveloped them just before the arrival of the Laurentide Ice Sheet, suggest that local ice in that district preceded the ice sheet and later merged with it. Evidence of such ice could be expected only in the stratigraphy of the glacial deposits, and must therefore await more detailed field study.

It is worthy of note that Chalmers (1895, p. 91M) inferred that the local ice was thin. Whether this ice preceded the main ice sheet, followed it, or both, thinness is expectable.

[3] For example, in a striation trending north-south he has indicated whether local evidence proves movement from south to north, or vice versa, rather than leaving the specific determination open.

6. Shickshock Mountains

The Shickshock Mountains, in the north-central part of the Gaspe Peninsula, have the character of a deeply dissected plateau. They form a belt about 15 miles wide, have a maximum altitude of 4160 feet, and center near 49°00′ N. Lat. and 66°00′ W. Long. The higher parts are shown on Map No. 2060 (Mount Albert Sheet) published by the Canada Geological Survey and on Maps Nos. 250 (Tabletop) and 111 (Lesseps) published by the Quebec Department of Mines. Well-developed cirques have floor altitudes of about 2500 feet.

After a field study of the higher areas Flint, Demorest, and Washburn (1942) reached these conclusions:

(1) Late in the Wisconsin age a local ice cap existed on the Shickshocks and flowed radially outward from them.[4]

(2) Later small glaciers flowed outward from cirques in the rims of the plateaulike higher summits. The late date of these glaciers is established by the presence of end moraines, related to the cirques, undisturbed by glaciation of a more general nature.

(3) That the Laurentide Ice Sheet overtopped the Shickshocks is possible but was not proved by the facts observed.

Alcock, who had earlier suspected (Alcock, 1935, p. 125) burial of the Shickshocks beneath the general ice sheet, subsequently established this relationship (Alcock, 1944). The principal evidence consists of numerous erratic boulders in the high country south of but lower than the Shickshock summits. The boulders, chiefly granite gneisses, were derived "clearly from the Precambrian country north of the St. Lawrence." Alcock also recorded a number of striations which he attributed to the Laurentide Ice Sheet.

7. "Appalachian" Center

Evidence, some definite and some only suggestive, of northward movement of glacier ice in southeastern Quebec has led to the hypothesis of a center of radial outflow in the Appalachian highlands of northern New Hampshire, extreme western Maine, and northern Vermont, where many summits have altitudes above 3000 feet. Although the former existence of such a center has not been conclusively established, evidence justifies retention of the hypothesis pending additional data.

The problem has two aspects: the existence of the supposed center, and its geologic date.

The hypothesis itself, and most of the supporting evidence, are the work of R. W. Ells and Robert Chalmers, of the Geological Survey of Canada. It was first proposed by Ells (1887, p. 45J), after two seasons' study of Compton, Stanstead, Beauce, Richmond, and Wolfe counties, Quebec. He found that a considerable proportion of the observed striations in which there was evidence of direction of flow had northwest bearings, although he stated that in many cases it was not possible to determine in which of the two possible directions the ice had moved. He found evidence of this type, indicating northward flow, as far west as Lake Memphremagog. In the vicinity of Lake Megantic Ells found granite boulders apparently derived from highlands to the southeast, although this provenance seems not to have been established conclusively. Ells' suggestion that local glaciers existed appears to have been influenced (Ells, 1887, p. 44J) also by the evidence of local glaciation in Gaspe and in New Brunswick, earlier reported by Chalmers and since confirmed.

After two seasons' additional field study in Megantic, Dorchester, Bellechasse, Levis, Montmagny, and l'Islet counties, Ells (1889, p. 98K–101K) listed additional striations embodying evidence of northwestward flow and concluded that the hypothesis of local glaciation appeared to be sustained.

Chalmers (1890) proposed the name *Appalachian system of glaciers* for the inferred ice mass. Soon afterward J. W. Dawson (1893, p. 150) showed radial ice flow from the northern Appalachians on a map.

Ells' preliminary field studies were followed by 3 years of more detailed field work by Chalmers, and, still later, by an additional 3 years' work. The resulting publications (Chalmers, 1899; 1906; 1908) adduce further evidence

[4] Additional evidence of this ice cap is the presence of boulders and of the distinctive Tabletop granite, derived from one of the summits, in valleys and on interfluves north as well as south of its outcrop area (F. F. Osborne, personal communication).

and summarize the case for local Appalachian glaciation. The evidence as a whole is of three kinds: (1) striations, (2) provenance of stones in the till, and (3) stratigraphy of the till.

(1) Chalmers listed the striations in southeastern Quebec and plotted them on a map (Chalmers, 1899, map following p. 160J) on which he differentiated striations (at 96 localities) recording northward flow from those (at 282 localities) recording generally southward flow. The former appear to be in addition to those listed (but not shown on maps) in the two publications by Ells. Chalmers laid emphasis on evidence of "stossing" in each locality of occurrence of striations. He concluded that this region had been covered both by local "Appalachian" ice and by the Laurentide Ice Sheet, although at different times. From the relative distinctness of the two groups of striations he inferred that the "Appalachian" ice was thinner than the Laurentide ice.

The striations adduced as evidence of local ice are confined to the northwestern and northern sectors of the area believed to have been glaciated locally. However, it is evident that both striations and erratics in the southeastern and southern sectors would record flow toward the southeast or south, regardless of whether the ice centered in a local source or a distant one. Hence it would be very difficult if not impossible to distinguish the effects of "Appalachian" ice from those of Laurentide ice in those sectors.

(2) Chalmers noted that many glacial erratics occur at points up to several miles north of the most northerly known outcrop from which they could have been derived, and at altitudes higher than those of the apparent sources.

(3) Chalmers (1908, p. 71) wrote:

> "Wherever traces of Appalachian and Laurentian glaciers are found together [in southeastern Quebec] the accompanying boulder-clay falls into a corresponding two-fold division; that produced by the former is always the underlying, and contains only boulder [*sic*] from local rocks. The overlying boulder-clay of the latter glacier on the other hand contains a considerable proportion of material derived from rocks belonging to the north side of the St. Lawrence ... and is found overlapping the older beds as far south as the Notre Dame mountains [of southeastern Quebec], and is, in the western part, carried farther south through the gaps and passes of the hills along the boundary line."

These facts he considered as evidence of local ice succeeded by Laurentide ice.

Chalmers dated this "Appalachian" center as earlier than the Laurentide Ice Sheet. Supporting evidence was found in the stratigraphic relation between the tills and in the presence, at four localities (Nos. 68, 92, 93, and 94), of northwest striations crossed by later south, southeast, or southwest striations. Evidently the early date of the local ice received support in Chalmers' mind from the deduction that the highlands of northern New Hampshire and adjacent Maine were

> "very favorably situated, geographically and meteorologically, on the advent of glacial conditions, for the production of glaciers. The greater amount of precipitation there compared with that of the region to the north of the St. Lawrence, and its position with respect to the Atlantic Ocean, combined to render it as suitable a gathering-ground for ice as Greenland is at the present day. It is not unreasonable to assume, therefore, that the ice began to form there independently, ... probably before it gathered upon the Laurentian plateau." (Chalmers, 1899, p. 40J).

The last statement fails to take account of northward decrease in temperature, which would offset the effect of precipitation, and the writer believes it goes too far. However, undoubtedly with suitable reduction of temperature these highlands both possessed an altitude and received an amount of precipitation that would favor the accumulation of glacier ice.

The principal criticism of the hypothesis of a local "Appalachian" center was made by MacKay (1921, p. 51–56). During three seasons of mapping of the Beauceville area, comprising 115 square miles, near the center of the region earlier examined by Ells and by Chalmers, he had found the concept of local glaciation to be at variance with the views of other geologists (unspecified) and in consequence devoted special attention to the evidence while in the field. His pertinent observations and conclusions were:

1. *Striations.* He recorded 655 striations and plotted them on an azimuth chart (p. 53). This revealed two predominantly preferred directions, the stronger one averaging N.14°W.-S.14°E. true, and the weaker one averaging N.66°W.-S.66°E. true. These directions are

clearly evident on Chalmers' (1899) map. MacKay also could not determine, in most cases, in which of two directions the ice was flowing. In the few localities in which he made this determination, he found evidence of apparent movement toward the southeast. Nevertheless on some bedrock bosses he found stossing on the south and southwest sides. He suggested that this condition might have been brought about by ice flowing southeast, but did not explain how this could have been accomplished.

2. *Till.* In a number of exposures the till consists of two members, an underlying unoxidized layer having a locally derived lithology and an overlying somewhat oxidized layer having a lithology of northern aspect. Both layers contain stones derived from points at least 10 miles to the north. MacKay was not certain whether these layers represent two distinct tills or the basal and superglacial drifts of a single glacier.

3. *Erratics.* Numerous erratics lie as much as a mile north of the bedrock localities from which they were apparently derived. At face value, this fact supports the northward flow of ice. MacKay, however, suggested that possibly (1) other outcrops of the same rock type may lie north of the erratics, but are concealed beneath drift, and (2) the erratics might have been moved northward along the beds of preglacial streams or on ice rafts in temporary glacial lakes.

These same two arguments, of course, can be applied to erratics apparently derived from near-by localities to the north against the theory of south-flowing ice. As used here, they suggest special pleading.

MacKay concluded that although the evidence does not all point in one direction, it favors the view that all the ice in southeastern Quebec came from the north. However, he added that

> "During the growth and southward advance of the Laurentian ice-sheet, local glaciers may have formed in the higher parts [of the Maine—New Hampshire—southeastern Quebec highlands] and have spread for varying distances down the mountain slopes until they finally coalesced with it, and the reverse order of conditions may have taken place during the closing stages of glaciation." (p. 56)

He ended with a mention of local cirques (with altitudes as low as 1000 feet) as evidencing local glaciers. Although it seems unlikely to the present writer that features as low as 1000 feet are actually cirques, A. L. Washburn (personal communication) reported that a cirque with a floor at 2500 to 2700 feet indents the southeast side of Jay Peak, Vermont (44°55′ N. Lat., 72°31′ W. Long.).

MacKay seems to have maintained an open-minded attitude, and his conclusions were nearly equivocal. We must bear in mind that the area he examined represents only a very small part of the region covered by Ells and Chalmers.

Long after the publication of MacKay's findings, clear evidence of northward flow of glacier ice has come from two localities in southeastern Quebec. At Thetford Mines, a boulder more than 5 feet in diameter has been moved at least 5 miles northeastward from its nearest possible source (Cooke, 1937). This information is based on the detailed investigation of a mining district and is very accurate. The other locality is at Bedford, a few miles northeast of the northeast end of Lake Champlain. Here evidence, from striations exposed in a quarry, of ice that flowed in a direction east of north is unimpeachable (Clark, 1937).

Additional evidence of an "Appalachian" center was adduced by Alcock (1948, p. 13), who wrote:

> "During the field seasons of 1944–46, inclusive, the writer took a large number of striation bearings in southwestern New Brunswick, both on the mainland and on the numerous islands, including Grand Manan. These coastal regions are heavily glaciated, showing well-developed *roches moutonnées*. The latter were produced by ice-sheets advancing from the land to the sea in directions ranging from southeasterly to easterly. The striation directions vary considerable [*sic*] but fall into several well-defined sets. Evidently, in addition to the advance of the Labrador ice mass, there was a movement from the mountains of Maine."

This is where the matter stands at present. That ice has flowed northward, at least locally, in southeastern Quebec is established. If Chalmers' evidence is accepted, this movement must have affected a sector at least 150 miles wide, measured in a southwest-northeast direction. According to the evidence, Chalmers'

inference that a center existed in northern New Hampshire and western Maine seems justified (with the addition of northern Vermont, in view of the western position of the Bedford locality) although more information is needed before this can be established. Even if it is later established, there will be great difficulty in fixing the southern limit of the ice, because the direction of flow would have been the same whether the ice was local or continental.

Chalmer's belief that the local ice antedated the main ice sheet was based on crossing sets of striations at four localities and on the presence of a till largely local beneath a till largely foreign. However, the great majority of Chalmers' occurrences exhibited only a single set of striations. As these features only rarely survive a second passage of ice over them, most of Chalmers' "local" striations probably postdate the general glaciation. Indeed, the evidence at Bedford (Clark, 1937) strongly supports the view that the north-flowing ice was the latest ice in that district. The stratigraphy of the till gives little support to Chalmers' dating, for, as MacKay showed, the lower till contains some foreign rock types. Furthermore, tills of predominantly local aspect, yet deposited by a continental ice sheet, are known in many parts of North America.

If, then, further facts confirm the hypothesis of an "Appalachian" center, it seems likely, as MacKay suggested, that this center both antedated and postdated the Laurentide Ice Sheet proper during the Wisconsin Glacial age. The rapid accumulation of a local ice cap may well have preceded the ice sheet, and the local ice may soon have been incorporated into the expanding continental ice body. Conversely during deglaciation, the high altitude of the northern Appalachian region, as compared with the St. Lawrence lowland northwest of it, may have favored the local persistence of ice upon the highlands while it was melting away from the lowland. This would have caused reversal of the direction of flow of the ice in the northwest sector of the highland district. This is apparently what happened slightly later when ice persisted on the Quebec-Labrador highlands after it had melted away, at least in part, from the Hudson Bay region.

Tarr, a profound student of existing glaciers in Alaska, favored the concept of local "Appalachian" ice, considered such centers expectable both before and after the main continental glaciation, and cited analogous conditions in the British Isles and in western Greenland. He believed that more than one "Appalachian" center may have existed (Tarr, 1900, p. 445–447).

The evidence at Bedford, Quebec, cited earlier, suggests a center in the Green Mountains in Vermont rather than one farther east. J. W. Goldthwait (1916b) reviewed early suggestions by Agassiz, Adams, Edward Hitchcock, and Charles Hitchcock that local glaciers had existed in northern Vermont. As a result of a brief field investigation he concluded that no valid evidence of *valley* glaciers in that region had been offered, and stated definitely that none of the higher summits (the altitude of the highest being 4393 feet) were characterized by cirques.[5] However, Goldthwait did not enter into the problem of a local ice cap, of which the evidence is necessarily very obscure. Detailed glacial study of northern Vermont is almost nonexistent; until such study has been made any conclusions as to the extent and number of "Appalachian" centers before and after the main continental glaciation must be held in abeyance. As yet we can say only that it is not unlikely that such a center existed in northern Vermont.

8. Upland of Southwestern Nova Scotia

The upland that constitutes the northwest part of the long southwest peninsula of Nova Scotia has been regarded as the site of a local center of radial outflow of ice during the Wisconsin Glacial age. This plateaulike upland lies in Kings and Annapolis counties, centering near 44°50′ N. Lat. and 64°50′ W. Long. Its general altitude ranges from 600 to 900 feet. The extreme northwestern edge of the upland, known as South Mountain, is underlain by a coarse gray granite. Along the coast of the Bay of Fundy, separated from South Mountain by the long Annapolis Valley, is North Mountain, a broad ridge of Triassic sandstones, basalts,

[5] This statement must now be modified by reference to the cirque in Jay Peak reported by Washburn.

and basic intrusive rocks, with a summit altitude of 700 feet.

The North Mountain volcanics are present on South Mountain, but boulders of the South Mountain granite are liberally present on the summit of North Mountain, several miles north of their apparent source (Dawson, 1893, p. 169). Coupled with the bearings of striations (shown on Pl. 1), they led L. W. Bailey (1898, p. 26M) to the opinion that an invasion by an ice sheet from the northwest was

> "followed by a period of more local and restricted distribution [of ice], when the higher portions of the peninsula became themselves the centre of the movement, the latter now occurring in all directions."

Faribault (1921, p. 8) corroborated this during detailed geologic mapping in the eastern part of this district:

> "Conclusive evidence was obtained of the general movement of the ice-sheet from north to south over North and South mountains and of the movement of local glaciers at the later stages of the Glacial period from South Mountain northward across North Mountain."

Walker and Parsons (1923) found that in another part of this district both striations and erratics of the granite (individually up to 12 feet in diameter) indicated glacier flow toward the northwest, at least through a short distance, at some time after a more vigorous general flow toward the south or southeast.

J. W. Goldthwait (1924, p. 81–82) concurred in the view that granite erratics are abundant on North Mountain. However, he suggested that they are not evidence of north-flowing ice, for two reasons. (1) some of the granites might have been derived not from South Mountain but rather from New Brunswick northwest of the Bay of Fundy. (2) Erratics identical with the South Mountain granite, "may have come from local patches of granite *north of the main* area, not yet discovered, and perhaps concealed by drift." The writer thinks Goldthwait's case seems like special pleading that fails to meet the statements of Dawson, Bailey, Faribault, and Walker and Parsons. Despite the low altitude of the district, the concept of a local center of radial outflow seems inherently probable. This part of Nova Scotia lies only 150 miles from the outer edge of the continental shelf, the probable outer limit reached by the Laurentide Ice Sheet. Thus it lies as close to the edge of the former ice sheet as do the Catskill Mounatins, but it lies 200 miles farther north, in a position favorable for the receipt of snowfall from frequent storms moving northeastward along the coast and along what was then the margin of the ice sheet. From inspection of the areal relations shown on the Glacial Map of North America (Flint and others, 1945) it seems probable that, when southwestern Nova Scotia was first uncovered by the waning ice sheet, Mt. Katahdin, the White Mountains, and the Adirondacks were still buried beneath the ice sheet, which at that early time was still thick.

9. Cobequid Mountains

The Cobequid Mountains are a highland mass southeast of the isthmus that connects Nova Scotia with New Brunswick. The approximate position of this mass is 45°30′ N. Lat. and 63° to 64° W. Long. The highest summit reaches 1020 feet, though most of the highland is considerably lower. The underlying rocks include syenite and felsite. North of the Cobequids is a lowland underlain by Paleozoic sedimentary rocks. Boulders of the sedimentary rocks occur on the Cobequids, but conversely the syenites and felsites are found as boulders on the lowland to the north (Dawson, 1893, p. 168; Chalmers, 1895, p. 53). From the Cobequids northward across the lowland to Northumberland Strait, Chalmers (1895, p. 96M, 103M; Map 559) found striations which he believed "stossed" at their southern ends, recording northward flow. He inferred that local ice had flowed outward from the Cobequids, toward the north as well as in other directions.

J. W. Goldthwait (1924, p. 83–84) recognized that a general ice sheet had flowed southward over the Cobequids. Likewise he confirmed the presence of Cobequid boulders north of their source area. He ascribed them, however, to northeastward flow of an ice sheet coming from New Brunswick. An examination of the directions of flow shown on the Glacial Map of North America suggests that Goldthwait's point of view is subject to doubt. Any flow from the New

Brunswick highlands across the Cobequids must have been south of east; how such flow could have transported Cobequid boulders northward is not apparent. In his discussion of Nova Scotia, the White Mountains, and the Green Mountains, Goldthwait seems to have started from the premise that cirques are a *sine qua non* of local glaciation. He seems not to have considered the possibility of local ice caps derived directly, by local reversal of flow, from the persisting nourishment of those portions of a thinning general ice sheet that lay over highland areas.

Wickenden (1941) identified three superposed tills, each lithologically distinct, near Joggins, Nova Scotia, and noted considerable variation in the directions of striations in the district northwest of the Cobequids. He did not infer the possibility of glacier flow from the Cobequids, but suggested that ice had flowed over this district in various directions successively.

10. Newfoundland

The entire west coast of Newfoundland consists of a broad highland (the Long Range) with an extreme altitude of 2600 feet. The south coastal area includes a highland whose summits, in the vicinity of Long. 56°40′, exceed 1500 feet. In contrast with these highlands, the central and eastern parts are much lower. Cirques at relatively low altitudes are present in the south coastal highland[6] and less commonly in the west coastal highland (MacClintock and Twenhofel, 1940, p. 1754). These cirques constitute evidence of small local glaciers. In addition, there exists evidence of radial outflow of local ice sheets. Striations in coastal areas are generally radial with respect to the island as a whole, suggesting an ice cap of island-wide dimensions. This general arrangement was recognized at an early date by Bell (1884, p. 37DD), who wrote: "On the island of Newfoundland, the glaciation appears to have been from the centre towards the sea on all sides." Also, more locally grouped radial striations suggest smaller ice caps in the Long Range and on the Avalon Peninsula in extreme southeastern Newfoundland. MacClintock and Twenhofel (1940, p. 1751) plotted the striations and two of the inferred centers. In the west-central interior, striations, stoss-and-lee topography, and the distribution of boulders of lead-zinc-copper sulphide ore in the drift suggest two former local ice caps in adjacent districts (Hans Lundberg, personal communication).

Finally, striations on the Port au Port Peninsula, in the southern part of the west coast, indicate southward flow followed by westward flow. Numerous large granite boulders on the broad surface of the highest west-coast summit, which is underlain entirely by quartzite, suggest glacier flow from Quebec and Labrador, on the northwest; such flow may have made the earlier, south-trending striations at Port au Port. Finally the very extensive evidence of consistent upwarping of the crust toward the northwest, along Newfoundland coasts, is strong evidence that the Laurentide Ice Sheet (Flint, 1940) invaded the island.

These facts and inferences suggest a sequence of events similar to that suggested for the Shickshock Mountains, though on a larger scale. During the Wisconsin Glacial age Newfoundland seems to have been invaded by the Laurentide Ice Sheet (although probably extensive local ice already occupied the island at the time of the invasion). Later, as the ice sheet thinned, radial flow from Newfoundland began, and still later (perhaps in sequence) the local ice cap separated into smaller and still more local caps. Finally, these gave way to cirque glaciers in restricted highland areas. Low temperatures and abundant precipitation, which still characterize much of Newfoundland today, favored persistence of these glaciers.

11. Highlands of Quebec and Western Labrador

The highlands of Quebec and western Labrador are essentially a hilly plateau sloping gradually down toward lower land in every direction and hence possessing an ill-defined boundary. These highlands extend, very roughly, from 62° to 72° W. Long. and from 51° to 55° N. Lat., with isolated high areas as far

[6] Observed by the writer in 1939. Well shown on Royal Canadian Air Force Photograph No. A 4590.9.

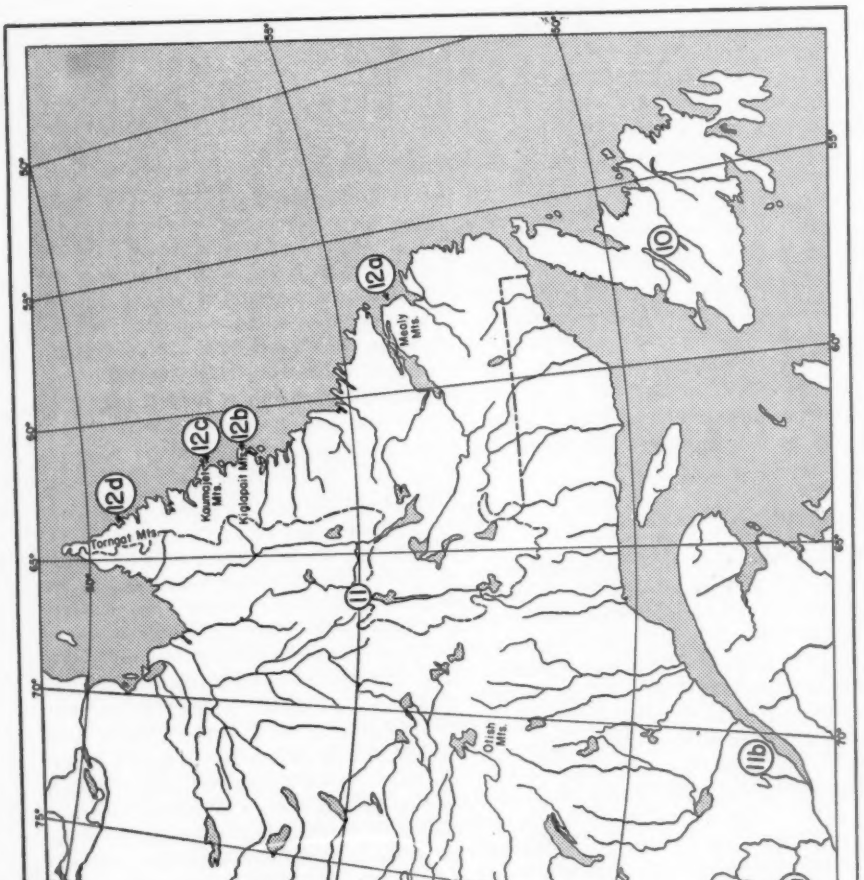


FIGURE 2.—EASTERN QUEBEC, LABRADOR, AND NEWFOUNDLAND
Showing known and probable highland centers of outflow of former glacier ice.
10.—Newfoundland 11, 11a, 11b—Highland of Quebec and western Labrador 12a, b, c, d—Coastal mountains of Labrador

south as 46° N. Lat. Most of their area lies at altitudes between 1500 and 3000 feet, although a few districts are higher. Of these, two are especially noteworthy, the Otish Mountains (3700 feet) approximately 52°20′ N. Lat.; 70°30′ W. Long.) and a highland (3900 feet) (at 47°40′ N. Lat.; 71° W. Long.) north of Quebec City. The other high points are close to the Gulf of St. Lawrence, between 66° and 68° W. Long. An extensive group of broad summits, locally reaching altitudes of more than 3000 feet, lies between 46° and 48° N. Lat., and 73°30′ and 76° W. Long.

This broad highland region as a whole was described by early explorers as covered with drift and sprinkled with boulders. It had been supposed, before exploration, that glacier ice had swept southward, from some vague source farther north, across the entire peninsula that separates Hudson Bay from the Labrador Sea. However, Bell (1884, p. 36DD–37DD; 1885, p. 14DD) reported evidence of the northward flow of former ice in the region between the highland and Hudson Strait, and Low (1893; 1902, p. 81D) recognized radial flow outward from centers on the peninsula. Thus the Quebec-western Labrador highland was early established as a conspicuous general center of outflow of ice within the Laurentide Ice Sheet (Chamberlin, 1894, p. 725, 731).

About the exact positions of specific centers far less is known. Low (1902, p. 81D) reported finding three distinct overlapping sets of striations, so placed as to suggest to him a progressively shifting center of outflow. He believed the earliest recorded position of the center was in the vicinity of 50° to 51° N. Lat., and 70° W. Long. The second position, in his view, lay to the northwest, near 54° N. Lat., whereas the third position lay near 55° to 56° N. Lat. and about 74°30′ W. Long. The inferred sequence implies a shift of 300 miles to the northwest.

Low appended to his report a list of the 145 striations, at 116 localities, on which his inferences were based. At only 21 of these localities did he record two of the three recognized sets of striations in superposition, and at only 3 localities did he report all three sets.

Statistically this evidence seems inadequate to the conclusion reached. Furthermore the geographic positions of the inferred centers deserve comment. The first or southeasternmost of his centers is very close to the Otish Mountains. Because of its relatively high altitude, this area seems a probable site of localized outflow near the beginning or near the end of a major glaciation. The other two inferred centers, however, occupy much lower ground. Even if the evidence furnished by striations were statistically satisfactory, a highland center would hardly be replaced by centers at successively lower altitudes toward the end of a glaciation—the only time to which a large number of existing striations are at all likely to refer. On both statistical and general grounds, therefore, the three successive centers inferred by Low may be fairly questioned until more compelling data appear.

Wheeler (1935, p. 245) inferred a "center of accumulation" of ice at about 57° N. Lat. and 70° W. Long., on low ground in the middle part of the Koksoak River basin. This location was determined by triangulation from striations on high summits in two widely separated areas near the Labrador coast. The inference appears to have been based on the assumption that, when the striations were made, the ice near the coast was flowing from a center far to the west rather than from the coastal mountains. Furthermore the striations and eskers along and northeast of the Kaniapiskau River, between the supposed "center of accumulation" and the Labrador coast, as compiled on the Glacial Map of North America (Flint and others, 1945), trend northerly, almost directly across the trend expectable on the basis of Wheeler's inference. Thus the inference of a center of outflow in the Koksoak River basin is believed not to be firmly based.

The data cited afford little basis for inferring the exact position of any center of radial outflow in the Quebec-Labrador highlands. Information from military air photography during World War II, however, added substantially to knowledge of the region. This information consists principally of the trends of large eskers; part of it is compiled on the Glacial Map of North America. The general arrangement of the eskers is radial with respect to a center on the height of land in the neighborhood of 55° N. Lat. and 67° W. Long., at an altitude of about 3000 feet.

10.—Newfoundland 11, 11a, 11b—Highland of Quebec and western Labrador 12a, b, c, d—Coastal mountains of Labrador
Showing known and probable highland centers of outflow of former glacier ice.

However, no esker information is available in the northwest quadrant of the region.

It is improbable that long eskers could have formed or could have been protected against serious modification by flowing ice before final deglaciation of the region. Hence the eskers on the Quebec-Labrador highland probably record the existence of a very late—perhaps the latest—center of outflow in the region. As the area central to the radial eskers is both high in altitude and central to the peninsula, it should have continued to receive a net accumulation of snowfall after other parts of the peninsula (excepting the coastal mountains of Labrador) had ceased to be areas of active accumulation.

In the southwestern quadrant of the peninsula, in the region southeast of James Bay, the eskers trend southwest and are paralleled by numerous striations. South of James Bay, on the other hand, eskers and striations trend southeast. Although no data on the relative ages of these features in the two regions seem to be available, the eskers and striations south of James Bay are probably the earlier, dating from a time when the wasting Laurentide Ice Sheet still covered not only Quebec but the James Bay region as well. The features southeast of James Bay probably date from a later time when the effective center of outflow of ice had shifted to the Quebec-Labrador highlands and when the ice in the James Bay region had thinned to a condition of stagnation or had disappeared.

Sharply localized radial outflow apparently occurred in some of the high districts immediately north of the St. Lawrence lowland. One of these is the highland (Figs. 1, 2, *11a*) developed on the Morin anorthosite and on granite, centering near 46°30′ N. Lat. and 74°30′ W. Long., northwest of Montreal, with an extreme altitude of 3100 feet. South of the highest area detailed field examination showed a set of east-west striations superposed on north-south striations, as well as boulders of the anorthosite at least 10 miles west of the westernmost known outcrop of that rock type. These facts suggested that the north-south striations record ice flowing from a center far to the north, whereas the east-west striations were made later, by ice flowing outward from a local center on the anorthosite highland itself (Osborne, 1938; supplemented by unpublished notes).

Somewhat similar though inconclusive evidence points to late outflow from a center (Figs. 1, 2, *11b*) on or north of the highland, 3900 feet in altitude, near 47°40′ N. Lat. and 71° W. Long.

F. F. Osborne has raised this interesting question: Did the existence of centers of relatively late date, such as these, north of the St. Lawrence lowland, delay the invasion of the lowland by sea water beyond the time when such invasion would have occurred according to a simpler hypothesis of deglaciation? If an "Appalachian" center of outflow existed southeast of the lowland, it too would have delayed the marine invasion. In this connection the suggestion of Antevs (1925, p. 66) is pertinent, that the ice in the Gaspe center may have been separated from the Quebec-Labrador ice by flotation as the sea began to invade the St. Lawrence lowland.

Detailed study of the areal geology of the Quebec highlands north of the St. Lawrence will probably produce conclusive evidence of other local centers that persisted late in the general history of deglaciation.

12. Coastal Mountains of Labrador

The coastal mountains of Labrador consist of a mass that was upheaved after having been reduced to vary low relief, and was then ruggedly sculptured. The upheaved mass is long and narrow, nearly paralleling the coast of the Labrador Sea. The mountains comprise four fairly distinct groups. From northwest to southeast these are:

(1) The Torngat Mountains, a continuous group 140 miles long, extending from the northern tip of Labrador at about 60°30′ N. Lat. southward to about 58°30′. The higher parts of the Torngats are a labyrinth of cirques and sharp peaks, whose highest summits exceed 5000 feet.

(2) The Kaumajet Mountains, a smaller group between latitudes 57° and 58° and reaching a maximum altitude of 4000 feet.

(3) The Kiglapait Mountains, between 56°30′ and 57° and having an extreme altitude of only 3000 feet.

(4) The Mealy Mountains, a compact group centering near 53°30′ N. Lat. and 59° W. Long. Their higher summits reach 4300 feet.

As an excellent description, review of the literature, and extensive bibliography of this region exists (Tanner, 1944; *see also* Odell, 1938) no summary of local glacial investigation is attempted here. Suffice it to say that near the coast, both in valleys and on summits, the direction of former glacier flow was northeasterly to the sea, and that no evidence of significant change in direction has yet been reported, with one important exception. The entire region was buried beneath the outflowing Laurentide Ice Sheet, and this condition of maximum glacial extent was followed and, in the Torngats at least, was preceded, by cirque glaciers (Tanner, 1944, p. 178–180). This sequence falls into a pattern already familiar from the facts and inferences from other highlands of northeastern America. Local glaciers formed on the coastal mountains were incorporated into the ice sheet encroaching from the west and, as the ice sheet waned, reappeared and were maintained by continuing snowfall at the lowered temperatures that still prevailed.

Along the coast such a sequence of events would have involved glacier flow in one direction only—northeastward to the sea. Evidence of reversal of flow should exist west of the coastal mountains, but, as that country is still virtually unexplored, we are ignorant of the glacial features it exhibits.

13. Eastern Canadian Arctic

The directions of flow of former glaciers in the eastern Canadian Arctic are so little known that we can only generalize about them. The information available up to 1942 is assembled, together with a full bibliography, in a comprehensive report by Washburn (1947, p. 44–60, 104–116, Pl. 3). Scattered supplementary data are given by Fortier (1948).

Presumably highland centers of outflow existed all along the mountainous highlands of Baffin Island (maximum altitude 10,000 feet), Bylot Island (6000 feet), Devon Island (7000 feet?), Ellesmere Island (9000 feet?), and Axel Heiberg Island (9000 feet?). The evidence of former glaciation is far too little known to establish this presumption, which is based solely on the fact that today each of these islands harbors one or more radially flowing highland ice caps, and on scanty local evidence that glaciers were formerly more extensive (Washburn, 1947, p. 57–58).

In addition, the glacial evidence suggests the possible existence of former centers at lower altitudes on Melville Peninsula, Victoria Island, Banks Island,[7] and Melville Island. However, far more information must be collected before more definite inferences as to the glaciation of this northern region can be made.

Conclusion

This review of the evidence presently available indicates that several highlands in northeastern North America stood high enough to maintain local glacier ice toward the end of the last glacial age, after they had ceased to be directly affected by the Laurentide Ice Sheet. In the White Mountains and in the Central Highlands of New Brunswick, evidence suggests that local glaciers not only postdated but also antedated the continental ice sheet. By inference, and by deduction from the meteorologic conditions that probably existed, it seems likely that on all or most of the other highlands discussed the ice sheet was preceded by local glaciers.

The facts and inferences yield a general reconstruction of glacial conditions more complex than has been generally realized. In the reconstruction the Laurentide Ice Sheet appears as the product of a coalescence of many local glaciers. Its disappearance is seen as having been complicated by the persistence of residual highland glaciers, some of which still exist.

Presumably the highlands described were the sites of glaciers because on them atmospheric temperatures were somewhat lower and precipitation was somewhat greater than on adjacent lower lands. It seems very probable, as has been shown by Leighly (1949, p. 138), that the first glaciers that formed in eastern North America at the beginning of any glacial

[7] Unpublished data supplied by A. L. Washburn from observations made in 1949.

age were those on the highlands along the lower St. Lawrence River. Leighly's compilation of present-day climatic data strongly supports this concept.

In addition to the highland centers of outflow there were also centers in areas of lower altitude. As the supposed meteorologic conditions in such areas are believed to have been unsuitable at all times for the accumulation of local glaciers, it has been argued (*e.g.* Flint, 1943) that at times during the glacial ages such areas, covered by the thick Laurentide Ice Sheet, were glacial domes—highlands of ice—which became sites of concentrated orographic precipitation. Thus they were meteorologically similar to the highland centers, although today they are topographically very difficult from them. Such lowland centers, of which there were probably many, do not form a part of the present discussion.

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HYDROTHERMAL-DIFFERENTIAL PRESSURE EQUIPMENT FOR EXPERIMENTAL STUDIES IN LOW-GRADE ROCK METAMORPHISM

By H. W. Fairbairn

Abstract

Details of equipment for combined hydrothermal-differential pressure experimentation are given. Fluid pressure up to 30,000 psi, compressive load up to 100,000 psi, and temperature up to 450°C have been successfully used. A considerable variety of experimental work can be attempted within these limits.

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Introduction

The value of experimental work in diagnosing geological problems needs no special pleading in these enlightened times. A great deal of significant work has been done, and much more remains. Most of the work up to now, however, has not utilized simultaneously all the pertinent geological variables—temperature, time, a chemically active fluid under controlled pressure, and an independent compressive load (with or without torque) on the specimen. Equipment which permits simultaneous application of all these variables was first used by Griggs about 10 years ago and resulted in a series of promising preliminary experiments (Griggs, 1941). The equipment he designed at that time is basically the same as that described here. Experience gained through more than 150 runs has resulted, however, in a number of modifications which greatly prolong the life of the equipment and improve its manipulation.

As no description of the original apparatus has ever been published, I wish to acknowledge Griggs' very essential part in its original design and his kind consent to publication of details of the equipment in its present form.

General Character of Equipment

The central feature of the apparatus is a vertical stainless-steel pressure chamber which sits on a hardened steel block (Pl. 1, fig. 1). A rectangular frame enclosing it rests on the piston at the top and at the bottom is connected to a horizontal bar. Weights on the end of the bar drive the piston into the top of the pressure chamber. Fluid (confining) pressure is maintained through a side connection in the upper end (Fig. 2, B). Heat is supplied by an insulated jacket which fits around the central part of the bomb. Temperature control is maintained by a thermocouple connected to an electric controller (Fig. 1). The experimental material is placed in the enlarged central area of the chamber and is housed in a receptacle of a design suitable to the type of experiment.

Details of Pressure Chamber

The pressure chamber is constructed of free-machining stainless steel, with a small-bore opening through the top to admit the piston assembly and a much larger opening through the bottom to admit the experimental material and the screw assembly which seals the inner

chamber. The flange near the top is bored horizontally to admit a connector supplying the confining pressure. The neck immediately below this flange is as narrow as is consistent with

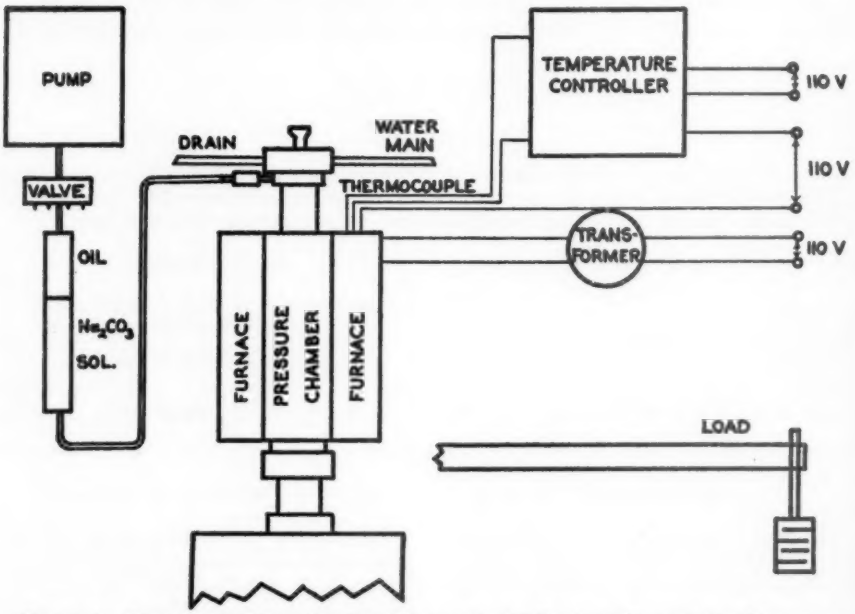

FIGURE 1.—DIAGRAMMATIC FLOW SHEET OF ELECTRICAL AND PRESSURE CONNECTIONS REQUIRED TO OPERATE THE EQUIPMENT

strength requirements in order to reduce heat losses. In the opening above the flange a shoulder is provided as a seat for a threaded piston assembly. In the lower part of the chamber another shoulder is provided as a buttress for the screw assembly. Two flats at the bottom aid in assembling and dismantling the appartus.

The platform (Fig. 2, D) which supports the experimental material is a solid stainless steel rod flanged at the top, extending from the shoulder in the lower part of the pressure chamber to the block on which the entire apparatus rests. To maintain the required confining pressure: (1) a knife-edged copper washer fits against the specimen platform, (2) an ordinary steel washer is behind this and (3) this in turn is backed by a hollow threaded cylinder with flats at its lower end (Fig. 2, E). Tightening this nut drives the knife edge of the copper washer into the space between the wall of the pressure chamber and the specimen platform, making a pressure seal which holds indefinitely within the temperature range used. To extract the experimental material after completion of a run, the base of the central stem or platform is provided with a left-hand thread and nut (Fig. 2, E). The procedure is first to loosen the large nut, then to tighten the smaller left-handed nut against the big nut. If the latter is then unscrewed, the copper washer seal will be broken and the entire assembly can be removed.

The piston assembly at the top is of hardened Ketos steel and consists of a nut (Fig. 2, A) threaded in its upper part and tapered below

PLATE 1.—VIEWS OF EXPERIMENTAL EQUIPMENT

FIGURE 1.—ASSEMBLED FOR OPERATION

Loaded bar extends from bottom left.

FIGURE 2.—PRESSURE CHAMBER (RIGHT) AND LOWER PRESSURE SEAL (LEFT)

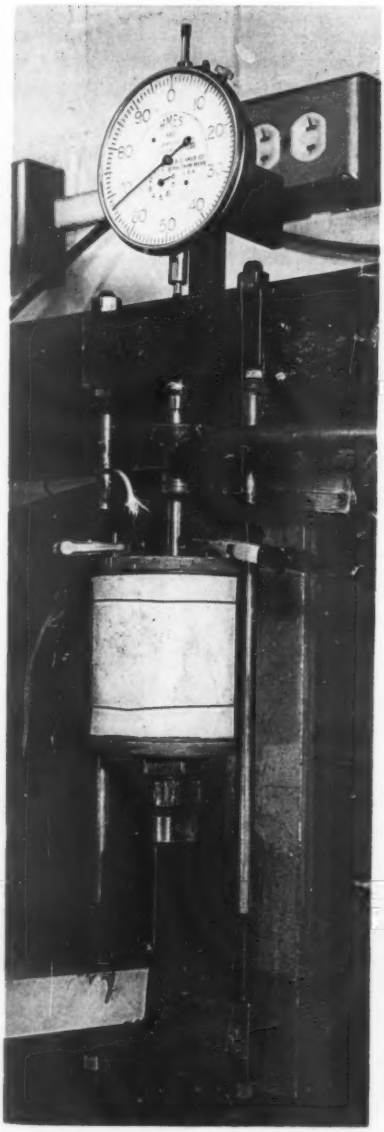

FIGURE 1

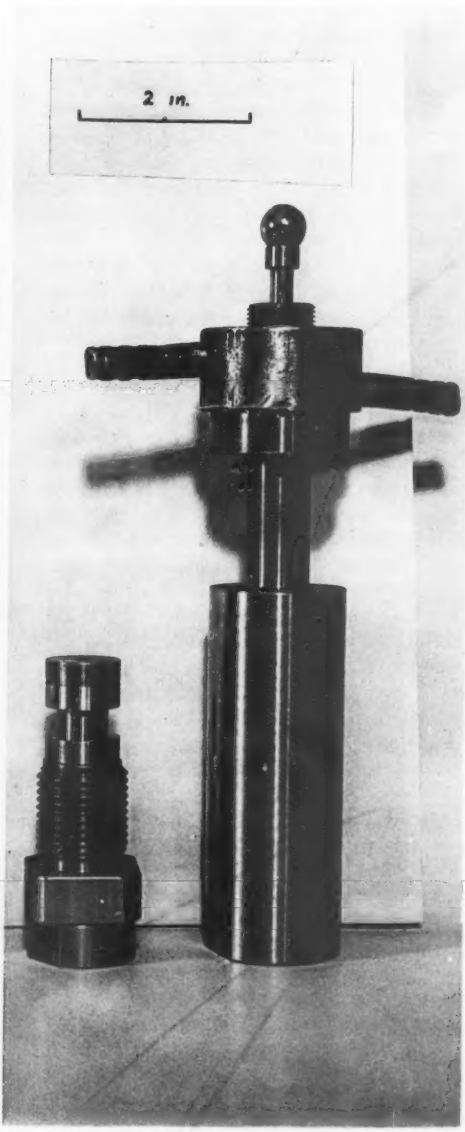

FIGURE 2

VIEWS OF EXPERIMENTAL EQUIPMENT

to make a pressure seal against the shoulder. The nut is fitted with a lapped, packless piston which under favorable circumstances will withstand as much as 30,000 psi without appreciable

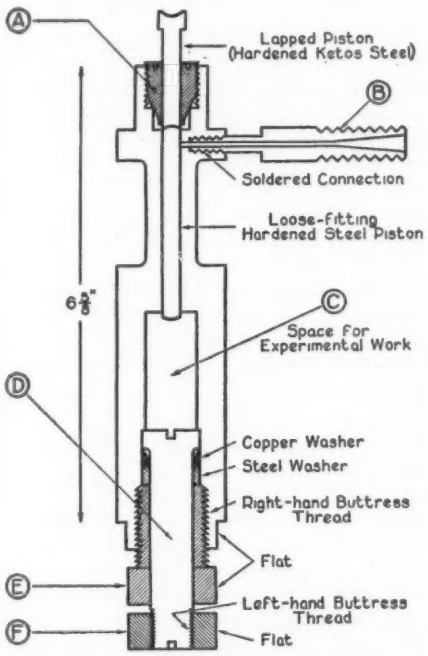

FIGURE 2.—ASSEMBLY DRAWING OF PRESSURE CHAMBER

leakage and yet slide with relative freedom. A steel ball connecting the top of the piston and the rectangular yoke enclosing the entire assembly (Pl. 1, fig. 1) provides a flexible union between the two units and maintains mechanical equilibrium. A rod of hardened drill steel (Fig. 2) connects the lapped piston with the receptacle containing the experimental material. It fits loosely within the pressure chamber and the rounded ends permit adjustment if alignment of the units is imperfect. This wide separation (4 inches) of the experimental chamber from the lapped piston is necessary so that the piston does not become hot and as a result unserviceable. It is cooled in addition by a water jacket (Pl. 1, fig. 2) just above the flange. The tapered nut (Fig. 2, A) is seated and extracted by means of a special spanner which fits around the piston and has lugs which fit snugly into the two holes in the top of the nut.

The horizontal connector (Fig. 2, B) in the side of the flange (through which the confining pressure is maintained) is threaded tightly into the flange and sealed with silver solder. (*See* Bridgman, 1931, Fig. 5, for details). The connecting nut (not shown in Fig. 2), with a left-hand thread at one end and a right-hand thread at the other, pulls the tapered ends together into a seal which can withstand at least 30,000 psi for long periods.

AUXILIARY UNITS

The yoke connecting the lapped piston to the loaded bar has threaded legs (Pl. 1, fig. 1) to permit adjustment for various experiments. The foot of a standard dial gauge rests on the top cross piece. A small screw clamp holds up the leg of the gauge while the experiment is being set up or disassembled. The lower cross piece of the yoke has a knife edge which fits into a V on the bar. A knife edge on the tip of the bar fits into a V on the under side of the block and acts as the fulcrum for the system. All these surfaces are hardened. The two contact points on the bar are spaced so that a mechanical advantage of 30 to 1 is obtained. Lead weights of 20- and 50-pound units, slotted and faced with smooth brass plates for ease in loading, provide the load on the other end of the bar. Standard 5-pound iron weights are also used, but lead for the larger units has a great advantage in economy of space and small length of hanger required.

The pump unit, tubing, and miscellaneous connectors, are standard items obtainable from the American Instrument Company.[1] The tubing, of molybdenum steel, is designed to withstand 90,000 psi. The pump is rated at 30,000 psi using a piston of 0.25 inch diameter. A distribution valve in the pump line allows five experiments to be run simultaneously. This type of multiple valve uses the unsupported-area pressure seal developed by Bridgman (1931, Fig. 8). Various combinations of gate valves and connectors of American Instrument Company manufacture can also be used for this

[1] Silver Spring, Md.

purpose. Between the distribution valve and any given pressure chamber, a special valve is needed to separate the oil on the pump side of the pressure line and the solution used on the pressure chamber side. This is accomplished in a thick-walled hollow steel cylinder having Bridgman pressure seals at each end and a "floating" valve in the internal chamber (Fig. 1). This valve consists of two steel washers enclosing a rubber washer, the latter fitting closely to the walls of the cylinder. At the time of loading it is placed in the extreme end of the cylinder and the chamber is filled with the required solution. To minimize mixing of the oil and solution, the cylinder is then mounted so that the "floating" valve is at the top and the oil is fed from the upper end. This separation chamber has a capacity of 0.5 liter and will function for a large number of runs without refilling.

The pressure chamber is heated by a removable cylindrical unit (Fig. 1; Pl. 1, fig. 1). The frame of this heating unit consists of a steel spool which is faced top and bottom with thick asbestos board to reduce heat losses. Two independent windings of Chromel resistance wire, isolated by thicknesses of asbestos paper, surround the spool. The inner is 26- gauge, 18.5 feet long, and is wired to an on-off type controller. The outer winding, 24-gauge, 23 feet long, is controlled by a Variac voltage regulator. The thermocouple fits into a small tube adjacent to the inner winding. In use, the Variac is set at that voltage which gives the closest control for the temperature desired, usually $\pm 4°$. For low temperatures, the outer winding is not required.

From the thermal standpoint the apparatus is highly inefficient, as unavoidable leaks at both ends of the pressure chamber disperse a considerable part of the heat energy. Nevertheless, a reasonably close temperature control is realized, as already noted. The actual temperature of the experimental material is not known from direct measurement, but can be obtained indirectly by calibration as follows. In the empty chamber, set on the block and connected as if for a run, a second thermocouple is inserted and connected to a potentiometer. For given temperature settings of the controller, the actual temperature of the center of the chamber is determined from the potentiometer geading. A graph is then constructed which gives the necessary controller setting for any desired temperature.

Durability of Pressure Connections

The pump, distribution valve, and separation valves rarely present any servicing problem; when leakage occurs, replacement of the rubber washers is a simple matter. The tapered coupling (Fig. 2) entering the pressure chamber (for maintenance of fluid pressure) will from time to time need refacing by an experienced mechanic. The copper seal in the lower part of the pressure chamber is almost infallibly reliable provided a new washer is used for each run. Difficulties which may arise are confined to extraction of the specimen platform and the assembly of washers and nuts beneath it. The steel washer backing the copper washer should be very loose-fitting and have tapered sides as a precaution against jamming. The use of buttress threads on the large locking nut reduces the probability of seizure, but it is advisable to be certain before a run that the threads are clean and smooth and that the nut fits rather loosely on both outside and inside wall. Jamming against the central stem (specimen platform) will sometimes occur if the latter gets out of alignment during an experiment. This progressive deterioration of metal subjected to long periods of heat and pressure can only be met by discarding parts long in use, preferably before mechanical trouble arises. The pressure chamber itself may bulge slightly long before actual failure occurs, thus serving notice of the condition of the metal.

The seal provided by the tapered nut at the top of the pressure chamber makes a reliable connection as long as the shoulder against which it rests has a sharp edge. With use it becomes beveled and must then be refaced. The lapped piston which slides within this nut is intended as a partial seal only and has a highly variable life. If fitted too loosely, the loss of liquid during a run (to maintain a given confining pressure) may prematurely empty the separation valve (Fig. 1). If fitted too tightly, the fraction of the compressive load delivered to the specimen becomes problematical because of increased friction. Moreover, the lapped surfaces not seldom deteriorate during the course of an

experiment, usually by tightening up. These parts of the equipment are the least satisfactory of all and a supply of spares must be kept on hand.

Range of Variables

The maximum fluid pressure safely used with this equipment has not been calculated. Present work has been confined to 30,000 psi, the rating of the pump. For any given experiment, moreover, the looseness of the lapped piston will control the practical upper limit of fluid pressure which can be maintained.

Temperatures higher than 450°C have not been used. The high and unavoidable thermal leakage through the top and bottom of the pressure chamber may set an upper limit to the attainable temperature of the experimental material. The calibration graph (controller setting vs. actual temperature) does not indicate where this might be, since within the range investigated the relation is a linear one only. It is certain also that the life of the stainless steel chamber will be considerably shortened in proportion to the magnitude of combined temperature and pressure.

The maximum load used on the bar is 150 pounds. Maximum safe load has not been estimated, but since 150 pounds produces slight curvature of the bar, loads have been kept within this limit. The actual load transmitted to the specimen is somewhat less than the 30-fold mechanical advantage set up by the fulcrum of the bar. The magnitude of this reduction, which arises from friction in the lapped piston, is unknown, but has been arbitrarily assigned a value of 10 per cent.

The only fluids used so far with this equipment have been water and alkaline solutions of various concentrations. These have no observable reaction with the stainless steel at elevated temperature and pressure. With acid solutions, the steel would in time undoubtedly become corroded. More critical, however, would be the reaction of an acid solution on the copper seal and the duration of an experiment at a given temperature and pressure would be entirely dependent on this factor.

Most of the experiments carried out with this equipment have lasted a few hours only. For longer periods (up to 117 days by actual trial), the quality of the lapped piston is the critical factor. If a loose piston can be tolerated, as in runs at low confining pressure, the piston will be serviceable for longer periods than where a tighter piston is required. Similarly, the pistons give longer service at low temperature than at high temperature.

Although the equipment has been used thus far only in experiments with monomineralic material and simple combinations such as quartz and calcite or quartz and clay (Griggs, 1941; Fairbairn, 1950), it seems likely that synthesis of more complex metamorphic fabrics could be accomplished within the range of variables indicated above.

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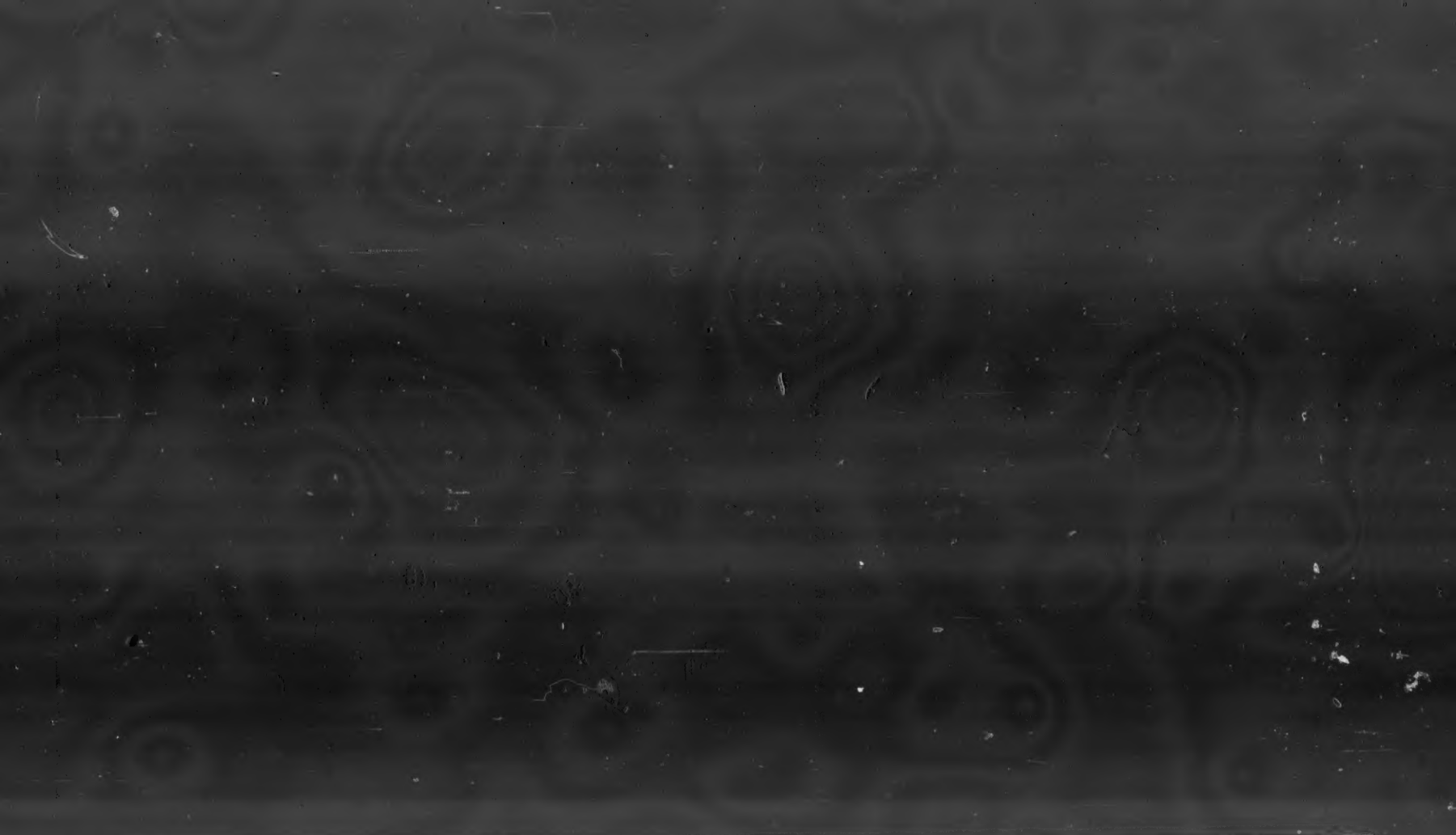

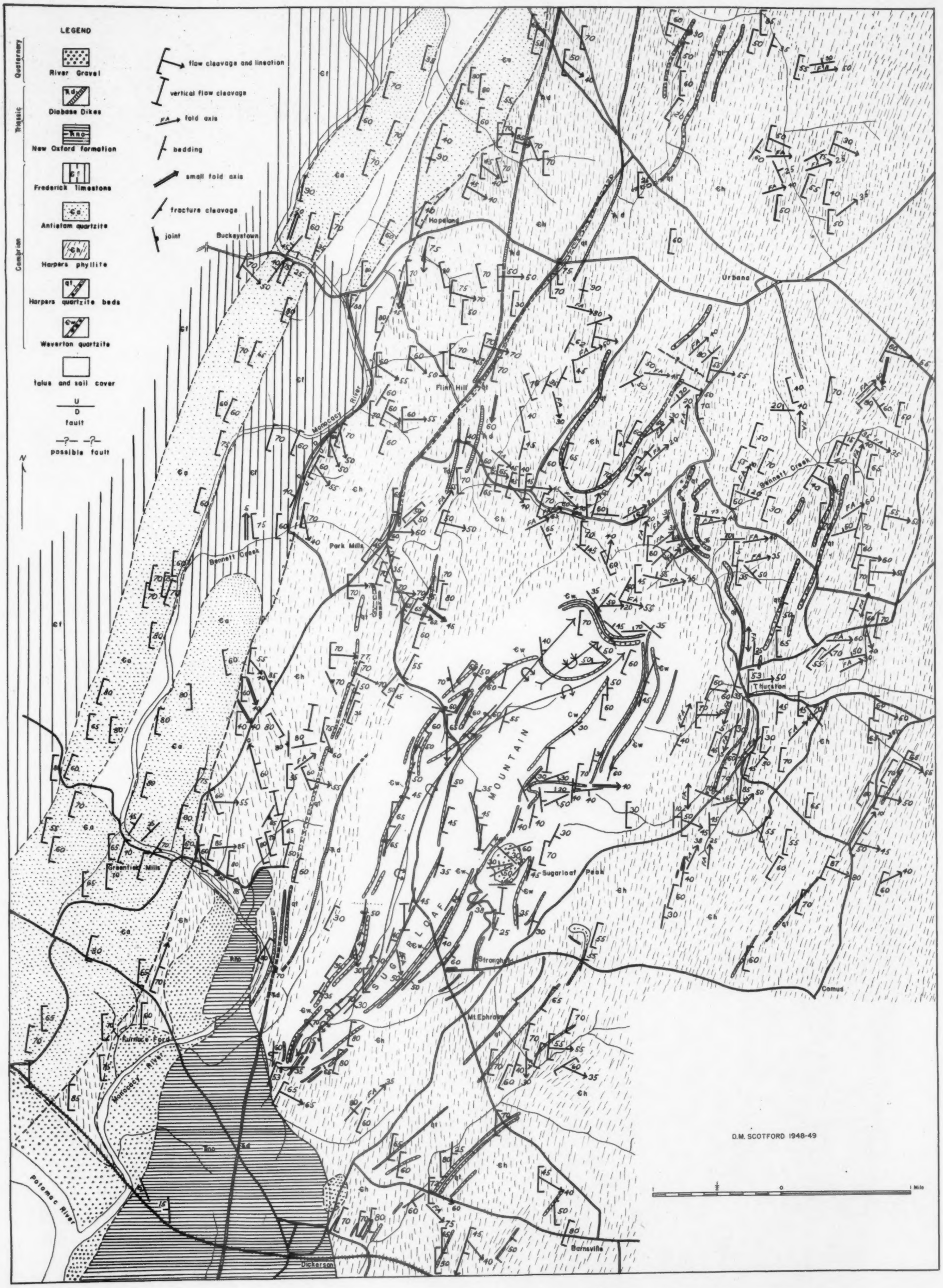

GEOLOGY AND STRUCTURE OF THE SUGARLOAF MOUNTAIN AREA, MARYLAND

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STRUCTURE OF THE SUGARLOAF MOUNTAIN AREA, MARYLAND, AS A KEY TO PIEDMONT STRATIGRAPHY

BY DAVID M. SCOTFORD

ABSTRACT

Detailed mapping of the bedding, cleavage, and lineation in the Sugarloaf Mountain area, Maryland, indicates the presence of an anticlinal dome. The structure, and consequent reversal of the stratigraphic sequence, and the lithologic similarity of the local quartzite and phyllites with those of known age 8 miles to the west indicate that the low-grade metamorphic rocks of this area are Lower Cambrian. It is hoped that this will shed light on the age of the highly crystalline metamorphics of the Maryland Piedmont. Some insight into the folding mechanism is afforded by the petrofabric study. The planes of liquid inclusion in quartz are apparently related to the *ac* cross jointing.

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INTRODUCTION

Location of the Area

The area is located 40 miles west of Baltimore and 10 miles south of Frederick, Maryland, on the border between the crystalline Maryland Piedmont and the Appalachian Province. The area mapped in detail (Pl. 1) comprises about 60 square miles.

Sugarloaf Mountain rises as a monadnock to 700 feet above the surrounding Piedmont upland. Two ridges form a horseshoe-shaped prominence open at the southwest end, 5½ miles long and 2 miles wide. The east ridge is broader and higher reaching 1282 feet at the south end.

Previous Work

The first detailed reference to the Sugarloaf Mountain area was by Keyes (1891), who interpreted the structure as a monocline with a possible double fault to account for the parallel ridges. Stose and Stose (1946) mapped it as an overturned, horseshoe-shaped syncline on one ridge and a normal syncline on the other. By this interpretation, the quartzite rests stratigraphically above the phyllite of the surrounding area.

Scope of the Present Study

This study is a detailed investigation of a key locality made in the hope that the result would shed light on the larger problem of the age of the Maryland Piedmont crystallines by indicating their relationship to the Appalachian sediments of known age just to the west (Fig. 1). Cloos (Cloos and Hietanen, 1941, p. 193) says:

"The author is fully convinced that the Glenarm series cannot be thrown into the Martinsburg group. The series is too little known and too complex for such a simple conclusion and it must include large portions of the Lower Cambrian sequence in a highly metamorphic facies. Further detailed and painstaking work will undoubtedly furnish the data on the basis of which more definite conclusions can be based. The answer seems to lie in the eastern Frederick County, South Mountain, Sugarloaf Mountain, and somewhere in the vast area which is now much too easily called Wissahickon schist and pre-Cambrian (?) because of 'lacking positive evidence to the contrary.'"

This investigation is an attempt at the detailed work suggested.

Method of Attack

The area was mapped at the scale of 1:12,500. Quartzite outcrops were traced along the strike, and formations plotted on the map only if actually exposed or clearly topographically expressed between outcrops. Elsewhere, the area is shown as covered. Hundreds of

measurements of the attitude of bedding, cleavage, lineation, and fold axes were made, but only representative symbols have been plotted on the map. The field work was done during the summer of 1948 and the spring and early summer of 1949.

One hundred twenty thin sections were studied. Forty-two petrofabric diagrams were prepared, and heavy mineral analyses were made of 35 quartzite specimens.

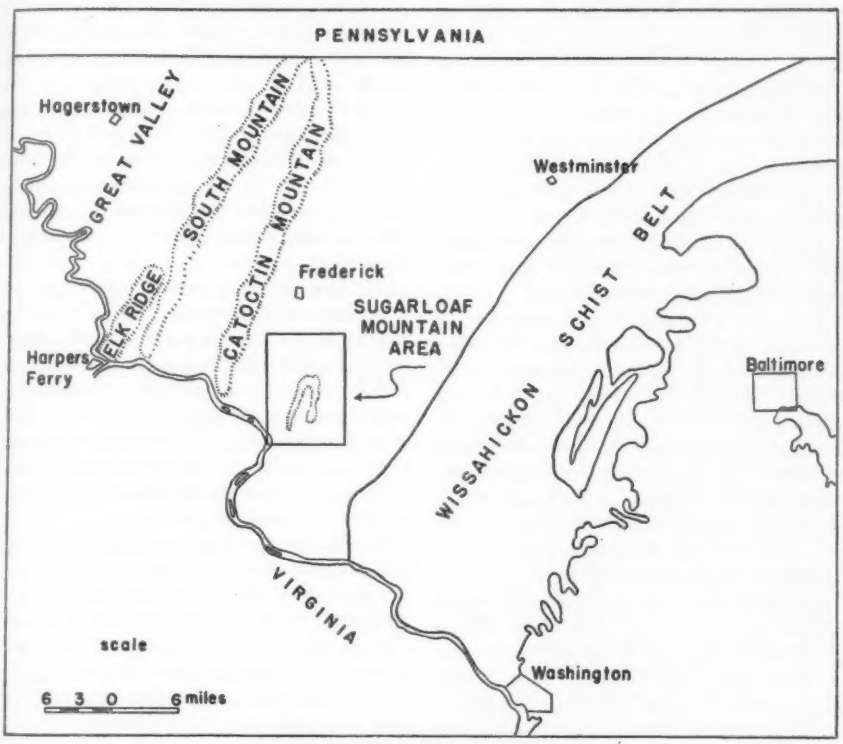

FIGURE 1.—INDEX MAP OF MARYLAND

Acknowledgments

The writer is indebted to Ernst Cloos for suggesting the problem and giving freely of his time for instruction and criticism. Thanks are also due J. L. Anderson for his assistance in the petrologic study. William Overstreet read the manuscript and made suggestions as to expression. Mr. Gordon Strong, owner of the southern portion of Sugarloaf Mountain, generously allowed the writer the use of a cabin on his land during the first summer. The writer wishes to thank his wife, Patricia T. Scotford, who assisted one summer in the field and did part of the drafting.

PETROLOGY

General Statement

Five major formations are exposed in the vicinity of Sugarloaf Mountain: the Triassic New Oxford formation, the Upper Cambrian Frederick limestone, and the Lower Cambrian Antietam quartzite, Harpers phyllite, and Weverton quartzite. A series of Triassic diabase dikes transect the area in a northeast-southwest direction about a mile west of the mountain. River gravels have been mapped on the west

side of the Monocacy River north of Furnace Ford and on the east and west banks south of Furnace Ford. A gravel deposit, possibly the Pleistocene (?) Lafayette formation, occurs on a hill just south of Mt. Ephraim.

Weverton Quartzite

General description.—The quartzites on Sugarloaf Mountain were considered Lower Cambrian and therefore Weverton by such early workers as J. P. Lesley, G. H. Williams (1892), and C. R. Keyes (1891). Jonas and Stose (1938) mapped these rocks as the Sugarloaf Mountain quartzites and assigned them to probable Lower Cambrian age but later (Stose and Stose, 1946) questioned the Lower Cambrian age.

The writer believes that these rocks are the equivalent of the Lower Cambrian Weverton quartzite because Sugarloaf Mountain is an anticline rather than a syncline, as previously reported (Stose and Stose, 1946). The stratigraphic sequence is therefore the reverse of that previously described and corresponds to the known Cambrian sequence in the South Mountain uplift. Phyllite lies above the quartzite just as on Catoctin Mountain, 8 miles west, Harpers phyllite lies above Weverton quartzite.

The Weverton is the principal ridge maker between the Hagerstown Valley (the "Great Valley") and Sugarloaf Mountain. From west to east it comprises Elk Ridge, South Mountain, Catoctin Mountain, and Sugarloaf Mountain (Fig. 1). It is considered Lower Cambrian because it conformably underlies the Harpers phyllite and the fossiliferous Antietam quartzite.

The Weverton on Sugarloaf Mountain is a massive, medium-grained rock varying on fresh surfaces from light tan to rusty brown, to maroon, to purple. The weathered rock is a rusty brown or dark maroon. The quartz grains are rounded and well packed in a siliceous groundmass. A few fracture surfaces are covered with small drusy quartz crystals. Quartz veins a few inches wide cut the rock parallel to some joint surfaces. Where visible, bedding appears as a color banding 2 to 6 inches wide, the darker bands being light maroon or brown (Pl. 2, fig. 1). At places the bedding is expressed as a fine dark-brown or black lining. Current bedding occurs in some layers made visible by this fine dark lining. It was not possible to establish the top and bottom of the beds on the basis of this current bedding alone. Fractures frequently parallel the bedding.

Cleavage is not prominent in the massive beds but is detectable on weathered surfaces by the parallel orientation of slightly elongated quartz grains. It has been confirmed by the petrofabric analysis of the included mica which lies in the cleavage. The rock is well jointed.

The Weverton quartzite beds on Sugarloaf Mountain (Pl. 1) are discontinuous lenses. Though some discontinuities are the result of faulting, most are thought to be depositional phenomena.

All areas on the mountain where quartzite is not exposed are covered by talus and other debris and have been so mapped. In the saddle separating Sugarloaf Peak from the next peak to the north, and on the western ridge of the mountain about 1 mile north of its south end, float of a dark-gray phyllitic rock was found. It is a quartz chlorite rock with much magnetite. Because of lack of exposures, its relationship to the quartzite is not known.

Microscopic characteristics.—Microscopically this quartzite may best be described in the manner used by R. E. Fellows (1943) for quartz tectonites, which contain five types of quartz: original grains, peripheral growth quartz, crush quartz, needle quartz, and recrystallized quartz. These quartz types will be described in the section on mineralogy.

In addition to quartz, there is usually a small percentage of short, tabular sericite crystals, which lie in the interstices between the original quartz. They are statistically parallel to the cleavage. There is usually a scattering of dust-like magnetite crystals, sometimes concentrated at the periphery of the original quartz grains. A small amount of detrital zircon and tourmaline is common, as well as hematite and leucoxene. Small quartz veinlets a few tenths of a millimeter wide cross the cleavage at various angles.

The original quartz grains are usually somewhat elongated. The average major axis is 0.8 mm, and the minor axis 0.6 or 0.7 mm. This elongation is statistically parallel to the rock cleavage. The grains are packed in a matrix of crush quartz having an average diameter of

about 0.05 mm. The abundance of crush quartz varies with the amount of deformation and ranges from 35 to 10 per cent of the rock, with the average being about 20 per cent.

If there is appreciable sericite as well as crush quartz in the matrix, the roundness of the original grains tends to be preserved. Under these conditions, peripheral growth aprons are commonly intact, and undulatory extinction, generally very common in the original grains, may be somewhat less prevalent. If peripheral growth aprons are present, the undulatory extinction usually includes them (Fellows, 1943).

Peripheral growth of quartz develops at an early stage of quartzite deformation (Fellows, 1943, p. 1415). Evidence for this is found in peripheral growth aprons which have been partially reduced to crush quartz (Pl. 4, fig. 2). Also peripheral growth is not common where the percentage of crush quartz is high; it has been removed during the deformation and has become crush quartz. The large original quartz grains, which are not common, are probably the first to be converted to crush quartz, because where they do occur they are highly fractured or have been completely reduced to crush quartz. The stress is then to a degree relieved, and the smaller original grains are not destroyed.

Heavy-mineral analysis.—A heavy-mineral study was made of 35 samples collected from quartzite on Sugarloaf Mountain and the Weverton quartzite exposed on Catoctin Mountain and South Mountain. The results were inconclusive. As shown by Pettijohn (1941), older sediments are generally poor in heavy-mineral varieties. The most abundant heavy nonopaque minerals at all localities were zircon, rutile, and tourmaline. Titanite and apatite were also present in minor amounts at all three areas. Anatase was present in the heavy residue from Catoctin Mountain but not from Sugarloaf or South Mountains. Black opaques, hematite, and leucoxene were very common in all locations. It is felt that the results indicate neither the presence nor the absence of a correlation.

Harpers Phyllite

General description.—The phyllites of the Sugarloaf area, which have been considered volcanic members of the Glenarm series (Stose and Stose, 1946), are the Harpers phyllite. This conclusion is based on the structural evidence, discussed below, that the local stratigraphic section is the reverse of that previously reported and therefore this section shows the same sequence of beds as the known Cambrian section 8 miles to the west. There is also a lithologic similarity, both megascopic and microscopic, between typical Harpers and the local phyllite.

Stose and Stose divided this phyllite into the Ijamsville and Urbana phyllites largely on the basis of color and the lack of calcareous bands in the Ijamsville. Although this division may be valid, the writer found it difficult to map the contacts accurately and considers them both to be part of the Harpers.

The Lower Cambrian age of the Harpers is based on its position conformably below the fossiliferous Antietam quartzite on Catoctin Mountain west of the Sugarloaf area (Stose and Stose, 1946). Except for Sugarloaf Mountain itself, the Harpers phyllite crops out over the entire area east of the Antietam quartzite belt. It is exposed along drainage, particularly Bennett Creek, and on many steep slopes. Where not actually exposed, its presence is readily indicated by abundant float.

The fresh phyllite varies from purple to light blue green to gray green, to dark green to almost black. The weathered rock may be brown or duller shades of the above colors. There is a continuous gradation from the pure argillaceous phyllite into the interbedded quartzite. The texture of the phyllite therefore varies from fine-grained to medium-grained, in which quartz grains become visible, to a coarse-grained rock which is almost pure quartzite. The black phyllite, which is exposed in a stream valley 1¼ miles south of Park Mills is fine-grained and composed largely of opaque carbonaceous matter with some chlorite bands and quartz.

The phyllite always shows a prominent cleavage. It is closely spaced in the argillaceous varieties becoming wider as the rock becomes more quartzitic. Bedding cannot be detected in most of the phyllite outcrops. Where visible it is usually a closely spaced color banding (Pl. 2, fig. 3). Occasionally secondary calcite has accumulated in the bedding plane giving

the rock a lighter banding (Pl. 2, fig. 2). In some places this calcite has been removed by weathering causing a honeycomb structure along the bedding planes.

Microscopic description.—Microscopically the phyllite is a strongly foliated quartz rock with varying amounts of chlorite which occurs both in the groundmass and as porphyroblasts. It is sometimes the only micaceous mineral. Albite is present in most specimens but always in minor amounts (less than 1 per cent). Rare bands of secondary calcite are usually parallel to the bedding. These bands are composed of medium-sized anhedral calcite crystals and quartz. Calcite also is occasionally disseminated through the rock in small amounts.

Zircon and tourmaline occur as accessory minerals, and a scattering of magnetite is common. Some of this may actually be ilmenite since a few specimens contain leucoxene. Hematite is found in many specimens either as anhedral crystals or as a stain on other minerals.

There is a complete gradation in the amount of quartz in the Harpers from a few per cent to over 90. Those beds with less than 65 per cent of quartz are considered phyllites, and those over 65 per cent quartzites. The average phyllite carries about 40 per cent of quartz as small to fine, angular to subangular grains which usually show some elongation parallel to the cleavage traces. A few grains show evidence of peripheral growth.

Sericite, the predominant micaceous mineral, reveals a strongly parallel orientation in the cleavage of the phyllite. In a few cases, sericite has been crinkled into a false cleavage (Pl. 4, fig. 3). Chlorite, in addition to its presence in the groundmass, also occurs in ovoid porphyroblasts with the longest direction parallel to the cleavage trace (Pl. 4, fig. 6).

The megascopic banding which denotes the bedding is seen microscopically to reflect variation in quartz content. The darker bands are more micaceous and less quartzose.

Metabasaltic bed.—The presence of a probable metabasaltic layer within the Harpers phyllite is indicated by a few outcrops three-fourths of a mile northwest of Barnesville. The rock is light green, fine-grained, and well foliated. It is composed chiefly of chlorite (50 per cent), quartz (20 per cent), and common epidote (10 per cent), with smaller amounts of magnetite, actinolite, and clinozoisite.

Light-green chlorite with its cleavage parallel to that of the rock occurs in the groundmass. Also blue chlorite is present in small irregular patches. Quartz makes up the rest of the groundmass, occurring as fine crystals associated with the green chlorite. Small to medium, tan, pleochroic, euhedral to subhedral, common epidote crystals are scattered through the rock. Clinozoisite looking much like epidote but with lower birefringence occurs in smaller amounts. Actinolite is present in about the same quantity. There is an irregular scattering of magnetite.

Quartzite beds.—The amount of quartz in the phyllite varies considerably, in some places increasing until it culminates in quartzite beds. As the phyllite is tightly folded and the quartzite beds are discontinuous along the strike, it is difficult to determine the number of quartzite horizons (Pl. 1). There are at least two principal layers and possibly more.

The quartzite forms ridges above the softer phyllite. A prominent ridge composed of quartzite beds and quartzitic phyllite extends from Park Mills northeastward for 5 miles. A less prominent ridge can be traced from Park Mills southwestward for 4 miles until it is covered by the Triassic arkose. A mile west of Flint Hill and a mile southwest of Urbana, a quartzite bed forming a U-shaped plunging syncline has been mapped. Several discontinuous quartzite beds striking generally 30° occur along the eastern border of the area (Pl. 1).

The average thickness of the quartzite beds is about 15–20 feet. The fresh massive rock is white, tan, or maroon. The more phyllitic quartzites are usually darker shades of brown, red, and maroon. The weathered rock is tan to brown or dark maroon. Since the quartzite, as well as the phyllite, contains a variable amount of quartz, the quartzite beds vary from a phyllitic or strongly cleaved quartzite to a massive pure quartzite with only a faint cleavage.

The phyllitic quartzite beds such as those northeast of Dickerson (Pl. 1) stand out above the interbedded phyllite but do not form as prominent ridges as the more massive layers.

The cleavage in the quartzite beds is always approximately parallel with that in the surrounding phyllite. Bedding has not been positively determined in the quartzite layers; however, since these rocks are interbedded members of a sedimentary formation, the strike of their bedding planes was determined by the direction of their contacts with the phyllite.

Microscopically the quartzite is composed of medium-sized, angular to subrounded, original quartz grains, in a matrix of fine-grained crush quartz and sericite; sericite lies in the cleavage. Small anhedral crystals of albite (Ab_{90-95}) are frequently present in minor amounts. A fine scattering of magnetite is common. Hematite stains are often present as a product of weathering. Minor amounts of chlorite frequently occur in small irregular patches. A few zircon crystals were found in some samples.

The original quartz grains are usually loosely packed in the crush quartz matrix. The degree of packing increases where the beds have been less deformed and less crush quartz has been generated. Some specimens contain original quartz grains with peripheral growth aprons. Some of these grains have partial aprons indicating that the rest of the peripheral growth has been reduced to crush quartz (Pl. 4, fig. 2). This suggests that the peripheral growth took place at an earlier stage than the crushing of the quartz grains.

The quartzite beds in the Harpers phyllite differ from the Weverton quartzite on Sugarloaf Mountain in several respects. The Weverton quartzite is cleaner, *i.e.*, it has no plagioclase and little sericite. In general the original quartz grains are more closely packed in the Weverton quartzite. In many cases rounded quartz grains are packed in a matrix of peripheral growth with little crush quartz present. Peripheral growth is less common in the Harpers quartzites. Tails of crush quartz and sericite (Fellows, 1943) are more common in the Harpers quartzites, as is the elongation of original quartz grains parallel to the cleavage trace. At a few locations the Harpers quartzites become massive and resemble the Weverton quartzite, but at most localities the characters of the two rocks are quite distinct. No evidence has been found of a gradation from the Harpers phyllite into the Weverton quartzite.

Correlation of the Harpers Phyllite

There is a close lithologic similarity both megascopic and microscopic between the phyllites in the Sugarloaf Mountain area and in the Harpers phyllite exposed farther west. This similarity, together with the identical stratigraphic sequence in the two areas, is the basis on which the local phyllite is correlated with the Harpers phyllite.

The rocks in both areas are essentially quartz-sericite phyllites with a variable amount of chlorite. In both areas, the phyllites contain chlorite porphyroblasts in the shape of elongated pods paralleling the cleavage trace.

In both areas, the quartz content of the phyllites varies markedly. On Catoctin Mountain, 8 miles west of Sugarloaf Mountain, two samples of Harpers phyllite were collected 200 feet apart. One sample contained 10 per cent of quartz, the other 95 per cent, with a complete gradation between the two.

Both areas contain quartzite layers interbedded with the phyllite. In a road cut on South Mountain near Rouzerville, Pennsylvania, E. Cloos has measured a section of Harpers phyllite containing two quartzite beds. In both locations, the bedding of the phyllite is exposed as a color banding usually at a large angle to the cleavage. In some samples from both areas, secondary calcite crystals are accumulated in the bedding planes. The quartz in both phyllites seems to be mostly authigenic and is generally elongated parallel to the cleavage trace. Zircon and tourmaline are common minor constituents of the phyllite in each area. Plagioclase of the same composition (Ab_{90-95}) is found in both phyllites. The gross mineralogy of the two phyllite areas is similar (Table 1).

The only difference in the two phyllites is the somewhat more common occurrence of chlorite, both in the groundmass and in porphyroblasts, in the Sugarloaf Mountain area (Pl. 3). This is to be expected as a function of the increasing grade of metamorphism of these argillaceous rocks toward the east.

Antietam Quartzite

The Antietam quartzite lies conformably above the Harpers phyllite on Catoctin Moun-

tain. It crops out in two southeast-trending belts, 1–2 miles wide, along the western portion of the Sugarloaf Mountain area. The western belt is a tight anticline which forms a ridge above the surrounding Frederick limestone. The eastern belt, which is discontinuous, is separated from the western by a limestone valley. The bedding in this belt reveals a series of tight folds where the Monocacy River cuts the Antietam east of Greenfield Mills.

The cleavage is strong. Commonly it has sheared out the bedding and makes difficult the determination of the attitude of the beds. In a road cut 1 mile north of Hopeland, a series of lenticular pods at a high angle to the cleavage express the vertical orientation of the bedding (Fig. 2). At other locations, the bedding is a faint narrow banding, but at most outcrops it is not determinable.

The Lower Cambrian age of the Antietam has been established on the basis of trilobite remains found on Catoctin Mountain 8 miles to the west (Stose and Stose, 1946).

The rock is a fine-grained arkosic quartzite. Fresh surfaces are grayish blue, and the weathered rock is rusty brown. Microscopically the Antietam quartzite is composed of uniform-sized, small, subangular quartz grains that make up 65 per cent of the rock. These grains are somewhat elongated parallel to the cleavage which consists of an alignment of fine-grained sericite that forms 10 per cent of the rock. Some of the quartz grains show strain shadows or undulatory extinction. A very fine-grained quartz makes up much of the groundmass. Small subhedral feldspar crystals, mostly twinned plagioclase (Ab_{90-95}) with some microcline constitute 20 per cent of the rock. Green tourmaline, chlorite, and magnetite are present in minor amounts.

Frederick Limestone

The Upper Cambrian Frederick Limestone lies above the Antietam quartzite at the western border of the area. A belt about a mile wide also occurs between Antietam ridges west of Hopeland to a point 4 miles southwest where it ends against Antietam quartzite. It is in fault contact with Harpers phyllite along the east bank of the Monacacy River a mile west of Flint Hill.

The Upper Cambrian age of the Frederick limestone is based on fossil evidence (Stose and Stose, 1946). The contact with the Antietam is probably unconformable, (Stose and Stose, 1946) since the Tomstown dolomite of Lower Cambrian age which overlies the Antietam to the west of the Triassic border is not present.

The rock is characteristically a dark-blue, thinly layered limestone with shaly partings separating the carbonate layers. Commonly shale predominates, giving the rock the character of a limestone conglomerate with the calcareous material occurring as pebbles in a shaly matrix. The color of the massive layers

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PLATE 2.—BEDDING IN WEVERTON QUARTZITE AND HARPERS PHYLLITE

FIGURE 1.—BEDDING IN WEVERTON QUARTZITE ON SUGARLOAF PEAK

FIGURE 2.—HARPERS PNYLLITE ON NORTH BANK OF BENNETT CREEK

Bedding parallel to knife. Flow cleavage dips steeply to the right.

FIGURE 3.—HARPERS PHYLLITE NORTH OF BENNETT CREEK

Bedding parallels the pencil. Flow cleavage dips steeply to the right. (Photographs by Joseph Schreiber.)

PLATE 3.—COMPARISON OF HARPERS PHYLLITE FROM SUGARLOAF MOUNTAIN AND SOUTH MOUNTAIN

FIGURE 1.—SPECIMEN 18, HARPERS PHYLLITE FROM SUGARLOAF MOUNTAIN AREA (LEFT). SPECIMEN 82, HARPERS PHYLLITE FROM SOUTH MOUNTAIN (RIGHT)

Sections cut normal to cleavage show similar bedding color bands. In Sugarloaf Mountain specimen, the bands are accentuated by the presence of fine porphyroblasts of chlorite, the grade of metamorphism being somewhat higher.

FIGURE 2.—SPECIMEN 18 LEFT AND SPECIMEN 82 RIGHT

Surfaces are parallel to cleavage. Similarity of bedding bands illustrated as in Figure 1. (Photographs by Spurbeck.)

FIGURE 1

FIGURE 2

FIGURE 3

BEDDING IN WEVERTON QUARTZITE AND HARPERS PHYLLITE

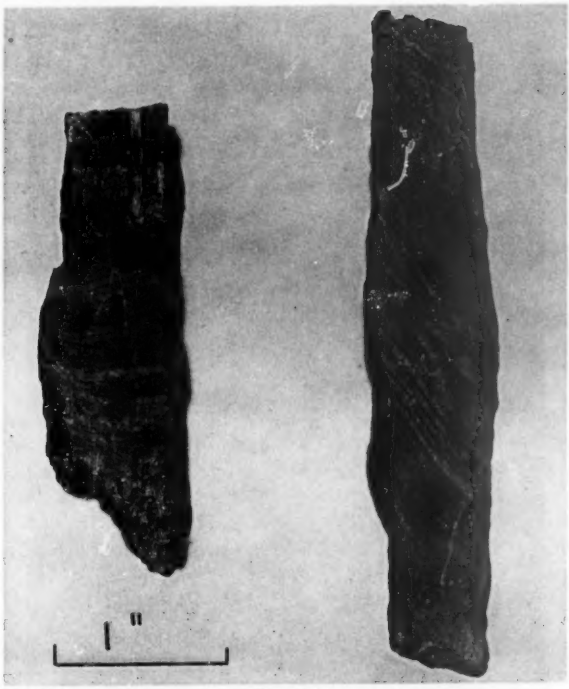

FIGURE 1

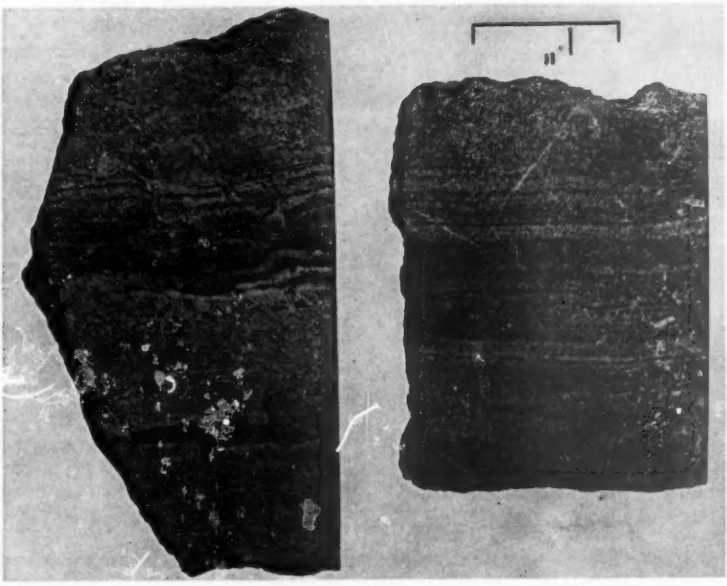

FIGURE 2

COMPARISON OF HARPERS PHYLLITE FROM SUGARLOAF MOUNTAIN AND SOUTH MOUNTAIN

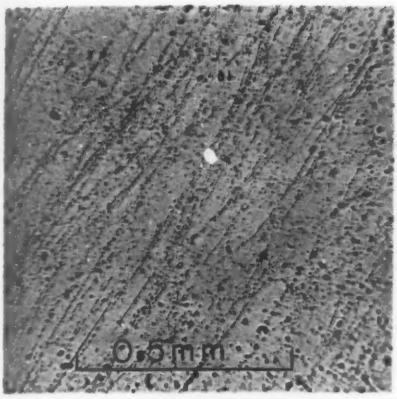

FIGURE 1

FIGURE 4

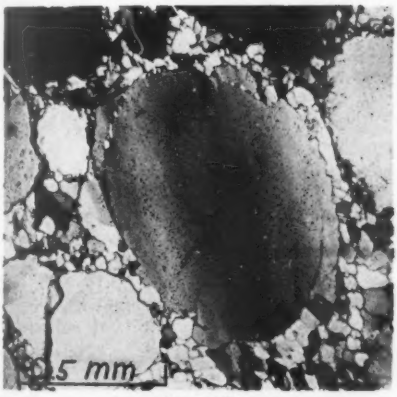

FIGURE 2

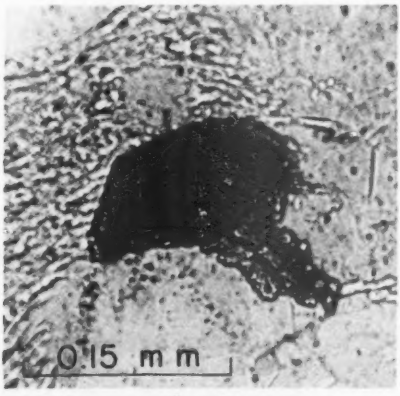

FIGURE 5

FIGURE 3

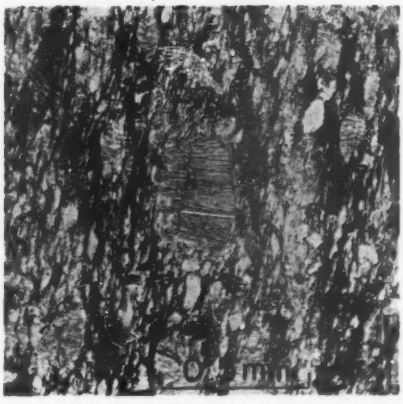

FIGURE 6

PHOTOMICROGRAPHS OF WEVERTON QUARTZITE, HARPERS PHYLLITE, AND FREDERICK LIMESTONE

FIGURE 1

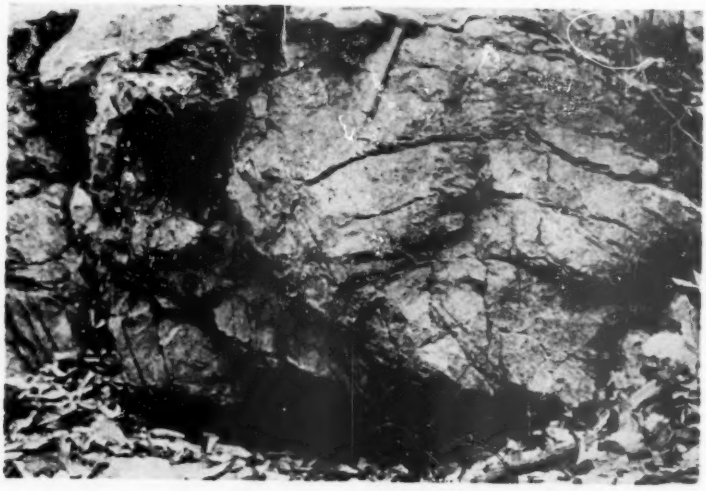

FIGURE 2

LINEATION IN HARPERS PHYLLITE AND DRAG FOLD IN WEVERTON QUARTZITE

varies from blue-gray to dark gray. The weathered rock is light gray.

In places the calcareous portions have been removed or pitted, leaving the argillaceous fraction in relief as bands of various thicknesses.

The rock has a moderately prominent cleavage.

Microscopically, the formation contains a large percentage of carbonaceous material, accounting for the frequent dark color. Some layers are oölitic (Pl. 4, fig. 4). Strongly deformed oölites of both the radial and concentric variety are elongated parallel to the cleavage. Some are radial at the center with concentric layers at the outside. The centers of the oölites usually contain small clusters of carbonate crystals. The matrix is largely fine-grained carbonate and carbonaceous matter. Small subangular quartz grains are scattered through the matrix. A few secondary calcite veins cross the cleavage.

New Oxford Formation

Since this formation had little bearing on the structural problem under consideration, its petrology was not studied. (For a discussion of the formation, *see* Stose and Stose, 1946, p. 84.)

In the Sugarloaf area, the rock is a dark maroon or brown, coarse conglomeratic arkose of Triassic age unconformably overlying the Cambrian Harpers phyllite. It occurs in the southwest portion of the area. The contact with the Harpers phyllite extends from the Monocacy River a mile north of Furnace Ford, southeast to Dickerson. The south nose of the Sugarloaf anticline lies just north of this contact. Fortunately erosion has removed the Triassic from this key location.

Mineralogy

General Statement

Over 20 minerals have been recognized. Quartz is by far the most common mineral with muscovite and chlorite also abundant. Zircon, tourmaline, plagioclase, magnetite, hematite, and calcite are common accessories. Microcline, orthoclase, epidote, clinozoisite, brucite, leucoxene, ilmenite, garnet, hornblende, and anatase are not common. Carbonaceous matter is present in some of the phyllites. As

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PLATE 4.—PHOTOMICROGRAPHS OF WEVERTON QUARTZITE, HARPERS PHYLLITE, AND FREDERICK LIMESTONE

FIGURE 1.—SPECIMEN 62, WEVERTON QUARTZITE

Crossed nicols. The traces of planes of liquid inclusions.

FIGURE 2.—SPECIMEN 70, WEVERTON QUARTZITE

Crossed nicols. Original quartz grain with peripheral growth apron partially reduced to crush quartz (lower right).

FIGURE 3.—SPECIMEN 122, HARPERS PHYLLITE

Parallel light. Fracture cleavage (vertical) crinkling flow cleavage (horizontal).

FIGURE 4.—SPECIMEN 118, FREDERICK LIMESTONE

Parallel light. Strongly deformed oölites.

FIGURE 5.—SPECIMEN 58, HARPERS PHYLLITE

Parallel light. Tourmaline with secondary growth appendage.

FIGURE 6.—SPECIMEN 18, HARPERS PHYLLITE

Parallel light. Porphyroblastic chlorite in matrix of sericite quartz and magnetite. Mineral cleavage of chlorite is normal (horizontal) to that of the rock (vertical).

PLATE 5.—LINEATION IN HARPERS PHYLLITE AND DRAG FOLD IN WEVERTON QUARTZITE

FIGURE 1.—SPECIMEN 108 HARPERS PHYLLITE FROM SUGARLOAF MOUNTAIN AREA

Showing strong lineation (vertical lines). (Photograph by Spurbeck.)

FIGURE 2.—DRAG FOLD IN WEVERTON QUARTZITE ON EAST LIMB OF SUGARLOAF ANTICLINE

(Photograph by Joseph Schreiber.)

TABLE 1. MODES IN THE HARPERS PHYLLITE

Specimen number	Quartz	Plagioclase	Microcline	Orthoclase	Carbonaceous matter	Sericite	Groundmass Chlorite	Porphyroblastic Chlorite	Magnetite	Zircon	Tourmaline	Calcite	Hematite	Apatite	Brucite
Sugarloaf Mountain Area															
2	60		T			30	8	2					T		
5	40						50		8	T			2		
6	60					38		2		T					
11	50	10				30		10		T					
15	60	5				30		T		T	T		2		
18	30	T				35		25							
16	45	T					30		5			10			
19	30	T				55		5	10						
20	20	T				55		20	5		T				
22	30					35			25		T		10		
32	20					65	T		10		T		5		
57	70	2				10	15		2	T		T			
77	20	T				15		20			T	45		T	
77a	60	T				20		10	10						
89	30						60	10		T	T				
90	20					30	30		20				T		
92	60	T						15	T	T		25			
94	20					60			20						
105	25	T				50			15		T	10			
114	50	2				35		5	10		T				
106	10				75	5	10								
122	30					65		5		T					
127	50	T				40	10		10	T					
Catoctin Mountain and South Mountain Area															
79	90	T	5			5			T	T	T		T		
82	30	T				70					T		T		
84	65	T				20	15		T	T	T				T
79G	30	T				25		3	2		T	30			
80G	75	5	T	T		20				T	T	T			
85	40	T				50	5		10	T	T				
86	85					15			T		T				
98	1					92		5	2				T		
100	65					35		T			T				
138	40	T				35		3	12	T	T				

T denotes less than 1 per cent.
Specimens 79G and 80G collected by J. E. Gair.

the grain size is invariably fine, it is frequently difficult to make optical measurements.

Quartz

Five types of quartz are present, original detrital quartz grains being the most common in the quartzitic rocks. Before deformation, they were fairly well sorted, round to subround, with an average diameter of 0.7 mm. Deformation and peripheral growth have made them more angular, somewhat elongated, and reduced many to crush quartz. Some grains have been completely recrystallized.

The original grains usually show undulatory extinction. Their borders are marked by dust particles, and they commonly contain liquid inclusions.

Peripheral growth as aprons of clear quartz around the original grains is common in many specimens.

Crush quartz, the small (.05 mm) subangular products of the deformation of original quartz and peripheral growth, fill the interstices.

Needle quartz is present but not common. These quartz slivers with a common elongation ratio of 3:1 are the products of early fracturing of original grains before rotation (Fellows, 1943).

Recrystallized quartz, the final product of quartz deformation, is common in varying amounts in all the rocks of the area. It is the product of the coalescing of crush quartz into larger elongated annealed grains (Fellows, 1943). This elongation is parallel to the cleavage trace of the rock, and is particularly common in the phyllitic rocks.

In places, tails of crush quartz and sericite are present at the ends of original grains caused by the coalescing of these two minerals. Syntectonic veinlets composed of recrystallized strained quartz are not uncommon. A few veins contain needle quartz and some post-tectonic veinlets are composed of undeformed clear quartz.

Muscovite

Muscovite, mostly of the fibrous sericite variety, is abundant in all formations. It lies in, and determines, the rock cleavage.

Muscovite is present in all grades of metamorphism, but the fine-grained sericite is a product of the lowest metamorphic grade. With increasing metamorphism the fibers become laths. Lathlike muscovite is not found in the rocks of the Sugarloaf area, but is common in the phyllites and schists a few miles to the east.

The sericite is frequently finely intergrown with chlorite and the two are closely associated with crush quartz.

Optical data could be obtained from only one specimen, because of the fine grain size. In this: $2V = 38°$; $n_\beta = 1.605$. This indicates an approximate composition of muscovite 38 per cent, Fe'''-muscovite 35 per cent, and phengite 28 per cent (Winchell, 1933, p. 268). This is similar to the muscovite found by Barth in Duchess County, New York, in which $2V = 30°–40°$ and $n_\beta = 1.600$.

Chlorite

Chlorite occurs predominantly in the Harpers phyllite but is also present in the Antietam quartzite, where it is gray green and pleochroic

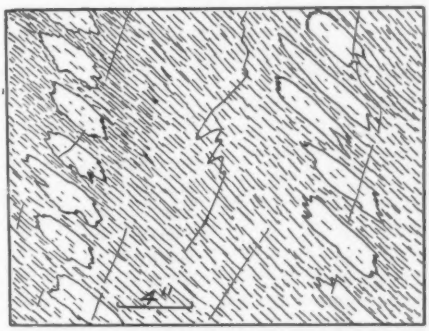


FIGURE 2.—CROSS-SECTIONAL SKETCH SHOWING PODS OF SHEARED-OUT BEDDING IN ANTIETAM QUARTZITE 1 MILE NORTH OF HOPELAND
The long axes of the pods parallel the cleavage while the trace of the bedding, which they reveal, crosses the cleavage at a large angle.

but too fine-grained and scarce for further optical work.

Two distinct types of chlorite occur in the Harpers phyllite: a light-green, somewhat pleochroic chlorite is intergrown with sericite in the groundmass of many phyllite specimens. This chlorite generally occurs in irregular patches scattered through the rock. Occasionally it comprises the entire groundmass, but is usually much less abundant (see Table 1). The other type of chlorite occurs in small oval or elongated porphyroblasts with the elongation parallel to the trace of the rock cleavage. The long axis is about 0.2 mm in both (Pl. 4, fig. 6). The cleavage of the chlorite is usually across the short direction of the porphyroblast, normal to the rock cleavage. These porphyroblasts were present in about half the phyllite specimens studied from the Sugarloaf Mountain area as well as locations farther west (Table 2). This chlorite shows a strong brown to green pleochroism and an anomalous dark-blue color under crossed nicols. The contrasting optical properties of the two chlorites are illustrated on Table 2.

Sixteen index of refraction determinations were made by oil immersion. In the case of the porphyroblastic chlorite, it was possible to obtain the value $(n_\gamma - n_\omega)$ by use of the Berek

Compensator after the mineral had been oriented in the universal stage to the position yielding maximum birefringence (with cleavage vertical).

ring as small subhedral twinned crystals dispersed through the rock.

To determine the composition of the plagioclase, extinction angles were measured on

TABLE 2. CHLORITE IN HARPERS PHYLLITE PORPHYROBLASTIC

Specimen number	Location	Sign of $2V$	n_β	$n_\gamma - n_\alpha$	Fe:Mg	Al:Si
18	Sugarloaf area	−	1.630	.011	3:2	1:4
77		−	1.633			
77a	Sugarloaf area	−	1.635	.006	3:2	3:7
89	Sugarloaf area	−	1.635	.006	3:2	3:7
92	Sugarloaf area	−	1.642	.004	7:3	1:1
110	Sugarloaf area	−	1.655	.004	4:1	3:2
84	South Mtn.	−	1.635	.005	3:2	2:3
85	South Mtn.	−	1.632	.007	3:2	3:7
G79	Harpers Ferry	−	1.631			
138		−	1.630			
98	Catoctin Mtn.	−	1.635	.004	3:2	2:3
	GROUNDMASS					
12	Sugarloaf area	−	1.615			
2		−	1.615			
16	Sugarloaf area	−	1.615			
57	Sugarloaf area	−	1.620			
5	Sugarloaf area	−	1.615			

The beta index for the porphyroblastic chlorite averages about 1.635 and is consistently higher than that for the groundmass type which ranges from 1.615 to 1.620. The porphyroblastic type, therefore, according to the data given by Winchell (1933, p.278), is probably iron rich with a uniformly high Fe:Mg ratio and tends to have a predominance of silica over alumina.

Although the value for $(n_\gamma - n_\alpha)$ could not be obtained for the groundmass chlorite, it can be inferred on the basis of the negative sign and lower index that this variety contains less iron, probably having a Fe:Mg ratio between 1:1 and 4:6. It must also be high in silica over alumina (Winchell, 1933, p. 278).

Feldspar

Albite is the most common feldspar and the only plagioclase present. It occurs in the Harpers phyllite and quartzite beds and in the Antietam quartzite. It is never more than 2 per cent of the rock and usually less than 1, occur-

the universal stage by the Rittmann (1929) method. Confirmatory determinations were made by the four-axis universal stage method of Turner (1947) and by oil immersions using the curves of Tsuboi (1923). The results were remarkably uniform, indicating the presence of an acid albite (Ab_{98-99}) in all the phyllite specimens investigated. This uniformity not only applies to the phyllites of the Sugarloaf area, but also to those specimens of phyllite containing plagioclase taken from Catoctin Mountain, South Mountain, and Harpers Ferry. A discussion of the significance of this uniformity will be reserved for the section on metamorphism. Microcline is present in small quantities in a few specimens of Harpers phyllite and is relatively abundant in the Antietam quartzite. It occurs as small subhedral crystals with plaid twinning. Orthoclase may be present in small amounts in the Antietam quartzite. However, the fine grain size makes determination difficult.

Zircon

Detrital zircon is a common accessory mineral in all the arenaceous and argillaceous rocks of the area. Small rounded to prismatic grains about 0.15 mm long are present in almost all specimens of these rocks. The heavy-mineral analysis of the quartzite revealed that zircon is by far the most common heavy-mineral constituent. The same is undoubtedly true of the phyllites.

Tourmaline

Tourmaline is a common accessory in the quartzites and phyllites. It occurs as small rounded grains about 0.1 mm in diameter or as small prisms or fragments. It is always far below 1 per cent of the rock. Its small size and scarcity made determination of its indices impossible, but the following pleochroic formulas were noted: E = yellow, O = blue; E = tan, O = brown; E = gray, O = gray-green.

The first suggests schorlite, the second dravite, and the third elbaite. In a few specimens irregular growth aprons of tourmaline, partially surrounding the original grains, were noted (Pl. 4, fig. 5). This evidently is a function of the low grade of metamorphism (Harker, 1932).

Although the tourmaline appears to the writer to be largely detrital, Barth (1936) sites the presence of tourmaline in low-grade metamorphic rocks as evidence of widespread pneumatolytic metamorphism.

Magnetite

Magnetite is a widespread minor constituent of all the formations. It usually is only 1 or 2 per cent of the rock but occasionally constitutes as much as 10 per cent. It occurs most commonly as a fine, well-scattered dusting throughout the rock, or as small irregular aggregates. The magnetite is frequently concentrated along cleavage planes or around original quartz grains. Occasionally well defined crystals as large as 0.15 mm are present in small quantities. A few thin sections contain a small amount of leucoxene in irregular patches, indicating the presence of ilmenite as well as magnetite.

Hematite

Hematite frequently stains many of the other minerals, particularly the sericite. It is also present as small irregular patches in the phyllite and quartzite.

Calcite

Subhedral calcite crystals constitute the largest proportion of the Frederick limestone. It is present in the oölitic layers in the centers of oöids as well as in the matrix. Calcite occasionally occurs as small anhedral to subhedral crystals disseminated in minor amounts through some of the phyllite specimens. In these it is commonly concentrated along bedding planes where it occurs as euhedral to subhedral crystals associated with quartz and sericite. Calcite is evidently the latest constituent of these rocks; it was introduced along the bedding planes after the folding. The crystals show no evidence of strain and occasionally appear to have replaced the sericite and quartz.

Epidote

Subhedral elongated epidote crystals 0.3 mm be 0.2 mm occur in the metabasaltic layer of the Harpers phyllite constituting about 10 per cent of the rock.

The optical data obtained were $2V = 70°$, $N_\alpha = 1.732$, $N_\beta = 1.766$. According to Winchell's (1933, p. 313) chart this indicates a composition of: 65% $HCa_2Al_3Si_3O_{13}$, 35% $HCa_2Fe_3Si_3O_{13}$. This is the common epidote, pistacite.

Clinozoisite

Clinozoisite constitutes about 5 per cent of the metabasaltic layer in the Harpers phyllite, occurs in rounded anhedral crystals about 0.2 mm wide, and looks much like epidote but has lower birefringence. Its $2V = 84°$, and it is biaxial positive. This indicates a composition of: 95% $HCa_2Al_3Si_3O_{13}$, 5% $HCa_2Fe_3Si_3O_{13}$ (Winchell, 1933, p. 311).

Minor Minerals

Minerals found only in the heavy residue from the Weverton quartzite were a colorless

detrital garnet, titanite, and a yellow anatase. These minerals constitute only a fraction of a per cent of the rock mass and are undoubtedly allochthonous.

METAMORPHISM

Introduction

A metamorphic terrane may best be described when the rocks are assigned to a mineral facies as defined by Eskola (1920, p. 146) or referred to a metamorphic zone after the manner first used by Barrow (1893).

The mineral facies or metamorphic zone will change with an increase in the grade of metamorphism. Thus a single geologic formation having a uniform chemical composition will have a changing mineralogy in response to the changing conditions of metamorphism to which it is subjected. The mineral assemblage, or facies, of the rock is then characteristic of the degree of metamorphism. Similarly the metamorphic zone changes with the changing grade of metamorphism, and a single mineral may be definitive of a given zone, the first appearance of that mineral marking the boundary of the zone.

Metamorphism in the Harpers Phyllite

Argillaceous sediments, through changing mineralogy, exhibit most clearly the successive stages of metamorphism (Harker, 1939, p. 208). Such a rock is the Harpers phyllite.

As previously mapped, the Harpers phyllite had its most eastern outcrops along Catoctin Mountain (Fig. 1) (Jonas and Stose, 1938). In this study its existence farther east in the Sugarloaf Mountain area has been recognized. Eastward from the Sugarloaf Mountain area, a belt of phyllites and schists, of increasing metamorphic grade, extends almost to Baltimore. This belt includes the crystalline Wissahickon and Peters Creek schists. While an investigation of this area has not been within the scope of the present study, it seems at least a distinct possibility that the Harpers is present in these crystalline rocks in higher metamorphic facies.

At Sugarloaf Mountain, the Harpers phyllite belongs to the greenschist facies (Eskola, 1939). Since the phyllite contains no biotite, but principally muscovite, chlorite, and quartz, it may further be defined as belonging to the muscovite-chlorite subfacies suggested by Turner (1948, p. 96). The state of metamorphism of these phyllites may be described also by the metamorphic zone to which they belong. The Harpers phyllite contains porphyroblasts of chlorite at Sugarloaf Mountain and as far west as Harpers Ferry. Thus this entire area lies within the chlorite zone and belongs to the greenschist facies.

These phyllites fit almost exactly Harker's (1939, p. 210) description of argillaceous rocks which have been metamorphosed chiefly through the action of shearing stress (dynamic morphism), but, unlike the lower grade within the chlorite zone, have also been affected by a certain elevation of temperature. Most of the detrital quartz has been recrystallized and occurs as lenticles with an orientation parallel to the schistosity. The tourmaline in some of the phyllites at Sugarloaf Mountain shows evidence of recrystallization. The hematite of the original sediment has been reduced to magnetite. The chlorite has been formed into patches or porphyroblasts with their cleavage transverse to that of the rock.

In all the specimens of phyllite of the Sugarloaf Mountain area, plagioclase, which occurs in small quantities, has the same composition—an acid albite (Ab_{98-99}). In those specimens of Harpers phyllite investigated from locations farther west, the same uniformity of plagioclase composition held true. Thus in the Sugarloaf area, and also as far west as Harpers Ferry (if the fewer specimens from more western localities can be taken as representative), the phyllite contains only a very acid plagioclase. It is difficult to determine if this is a detrital or an authigenic plagioclase. Possibly the source of this albite could have been a rock containing only this type of plagioclase from which detritus was spread over a wide area. It seems more likely, however, that the mineral is the product of the widespread low-grade metamorphism which affected the area.

Turner (1933) points out the increasing anorthite content of the plagioclase with increasing metamorphic grade in certain schists of South Westland, New Zealand. In the chlorite zone,

the plagioclase is albite with oligoclase entering at the almandite or kyanite zone. Hietanen (Cloos and Hietanen, 1941) found albite in the biotite zone of the Wissahickon schist of York County, Pennsylvania. In the almandite schist facies, however, oligoclase (Ab_{78}) is present, although plagioclase ranging from Ab_{81} to Ab_{70} was also found within the zone.

Harker (1939) recognized albite-chlorite-sericite schists which are represented in low grade metamorphism. He considers albite as typical a stress-mineral as the associated chlorite and sericite.

Metabasaltic Layer

The green phyllitic layer inclosed in the Harpers phyllite west of Barnesville, composed of chlorite, epidote, clinozoisite, actinolite, and fine-grained quartz, has been called metabasalt following the lead of Jonas and Stose (1938). However, in the specimens studied, there were no amygdules and no hornblende as described by them, though they may be present elsewhere. Jonas and Stose believed these layers to be amygdaloidal flows. It may be that these are represented here by a more tuffaceous phase.

Possibly retrogressive metamorphism has caused a reduction in the grade, and the original amygdaloidal flow mineralogy has been altered to that of the greenschist facies. Some evidence for this is found in the thin sections where chlorite appears to have formed pseudomorphs after hornblende.

Another possible interpretation is that these layers may not have been flows but argillaceous layers with a somewhat different original composition than the rest of the phyllite. This may have allowed the formation of epidote minerals (which are not found in the other phyllites) during metamorphism. Epidote minerals may also occur in the greenschist facies when the chemical composition of the rock allows it (Eskola, 1939).

Metamorphism in the Quartzites and Limestone

There is no change in the mineralogy of the Weverton quartzite or quartzite beds in the Harpers caused by metamorphism. In the Antietam quartzite, albite (Ab_{94}) is present as well as chlorite of the same composition as found in the groundmass of the Harpers phyllite (Table 2). The calcite in the limestone has been recrystallized, but no new minerals have been generated.

Megascopic Structure

Introduction

The data are here presented on which the conclusions regarding the structural configuration and stratigraphic sequence are based. An attempt is made to point out the relationship of the smaller features to the larger structural picture.

The terms flow cleavage and fracture cleavage are used as defined by Leith (1905; 1923). The geographic orientations of structural elements are read from a 360-degree scale with north at 0°. An east-west strike is thus simply called a strike of 90°, the smaller number being used in each case. Some features are presented in statistical diagrams (Figs. 3, 4). These diagrams are the lower half of equal-area net projections oriented into geographic position.

Bedding

Bedding has been described in the section on Petrology. While measurement of bedding attitudes is not possible at most outcrops, it is not so rare as to prevent a fairly clear structural interpretation.

Bedding is more prominent in the Weverton quartzite and Harpers phyllite to the east of Sugarloaf Mountain than to the west. This is probably caused by shearing out of bedding along the cleavage planes in the west. Since the bedding in the quartzite on the west ridge parallels the cleavage, its determination is usually difficult.

Cleavage

General statement.—Deformation has produced both flow and fracture cleavage, although the latter is of minor importance. Flow cleavage results from the recrystallization and reorientation of platy minerals; fracture cleavage is due to rupture along shear planes. A complete gradation between the two types is possible (Cloos and Hietenen, 1941, p. 24), but this does

not apply in the Sugarloaf area. Flow cleavage predominates in both argillaceous and arenaceous beds. This fact supports the conclusion,

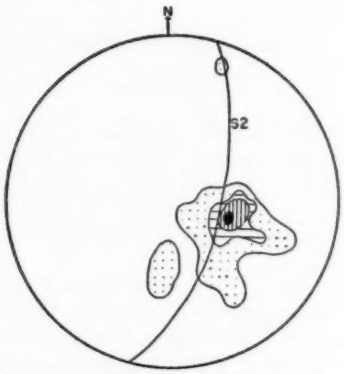


FIGURE 3.—STEREOGRAPHIC PROJECTION OF LINEATION
Maxima 26 per cent.

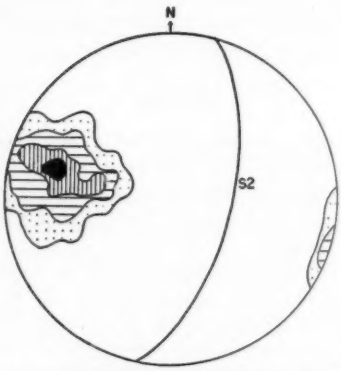


FIGURE 4.—STEREOGRAPHIC PROJECTION OF POLES OF FLOW CLEAVAGE
Maxima 2 per cent.

indicated by other evidence that flow was a prominent mechanism in the deformation of the area.

Flow cleavage.—All the rocks except the Triassic arkose exhibit a flow cleavage (schistocity, slaty cleavage). It is the most prominent structure in the phyllite, becoming less conspicuous only in the most quartzitic layers. Even here the parallel orientation of sericite is visible megascopically. In the Weverton quartzite it is revealed by a parallel elongation of the quartz grains. This cleavage is exhibited in the Antietam quartzite and Frederick limestone by a rather faint platy structure.

Microscopically the cleavage is always easily detected. In the phyllites the parallelism of the micaceous minerals produces the cleavage. The elongation of quartz grains and statistically parallel orientation of sericite reveal the cleavage planes in the Weverton quartzite. The elongation of oölites is the most obvious microscopic indication of the flow cleavage in the limestone.

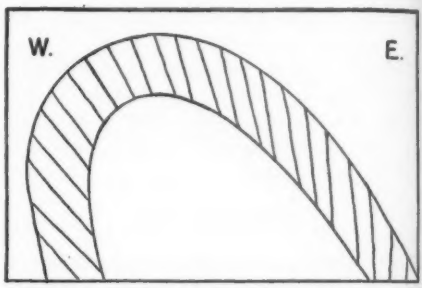


FIGURE 5.—ANGULAR RELATION OF CLEAVAGE TO BEDDING IN AN OVERTURNED ANTICLINE

The homogeneity of the orientation of flow cleavage is remarkable and probably a product of Appalachian deformation which affected a large portion of the Piedmont and Appalachian provinces. Figure 4 shows a stereographic projection of flow-cleavage orientation. The average cleavage, based on 100 representative measurements, strikes 17° and dips 60° E. This is represented by the 14 per cent maxima on the diagram. The strike varies from 340° to 60°, and the dip from 20° to 90°.

The abnormal western strikes are most common at the western border of the Harpers phyllite. They may be associated with the possible faulting at the contact of the Harpers and Antietam.

There is some indication that the cleavage tends to be steeper in the quartzites than in the phyllite. Many instances of vertical cleavage have been noted in the Weverton quartzite.

Relation of flow cleavage to bedding.—The angular relationship between flow cleavage and bedding on either limb of the fold is one of the important indications of the anticlinal nature of Sugarloaf Mountain (Fig. 5).

In the center of the fold the cleavage is parallel to the axial plane but fans slightly in the limbs. In the west limb of an overturned anticline, the bedding dips more steeply than the cleavage, while in the east limb the cleavage dips more steeply than the bedding (Nevin, 1942, p. 155).

In the east ridge of Sugarloaf Mountain the cleavage dips more steeply and in the west ridge less steeply than, or parallel to, the bedding, indicating the east and west limbs of an overturned anticline (Pl. 1).

Fracture cleavage (false cleavage, slip cleavage).—A closely spaced fracture system frequently parallels and accentuates the flow-cleavage planes especially in the weathered quartzites. This is not considered a separate fracture cleavage, but the product of later parting along earlier-formed planes of discontinuity.

Fracture cleavage as distinguished here transects all earlier structural elements at varying angles. It is not as prominent as flow cleavage but is important in a few locations particularly in the phyllite. It is expressed by either a closely spaced fracture or a crenulation of the flow cleavage planes (Cloos and Hietenen, 1941, p. 25). Frequently this crenulation is visible microscopically where it is not seen megascopically (Pl. 4, fig. 3).

Although only few measurements of this cleavage could be made, it does not seem to parallel the axial planes of the small or large folds. The strike is consistently more westerly than that of the flow cleavage, varying from 150° to 170°. The dip is always steep and usually to the east.

Lineation

Introduction.—Sander's (1930, p. 119) coordinate system is used: *b* is the fold axis; *a* is perpendicular to *b* in the movement plane; and *c* is perpendicular to *ab*. Fold axes will be discussed later although they are also a lineation.

Lineation in the Harpers phyllite.—Lineation in the Harpers phyllite becomes visible due to a crinkling on the flow cleavage surface (Pl. 5, fig. 1) and is frequently the result of the intersection of flow and fracture cleavage. Two such lineations have been observed.

The most prominent one lies in *a* or *a'* (Fig. 3). As can be seen, it is in the flow-cleavage plane (*ab* plane). A maximum of 26 per cent, indicating the highest concentration of lineation, lies close to the trace of the flow cleavage and strikes 105° and plunges 60°. This lineation is not only the most prevalent in the Sugarloaf area but is common as far west as South Mountain (E. Cloos, 1947, Pl. 10).

An abnormal, more southerly, orientation of this lineation is found along the western border of the Harpers phyllite near the contact with the Antietam quartzite and Frederick limestone. The lower 4 per cent maximum in the projection (Fig. 3) illustrates this orientation. The average strike is 170° and the plunge 55°. This abnormality may be related to the possible faulting at the contact and will be discussed later.

A second lineation is much less pronounced. It lies in *b* which is usually subparallel to the strike of the flow cleavage but varies in orientation north of the Sugarloaf anticline. The upper 4 per cent maximum (Fig. 3) expresses its orientation: strike 20°, plunge 10°.

Fold Axes

Two types of fold axes have been measured. The axes of small folds a few feet in amplitude which occur predominantly in the phyllite and to a lesser extent in the quartzite can be measured directly. The axes of the larger folds in the Harpers phyllite, which are not exposed in their entirety, are determined by the orientation of the lineation produced by the intersection of flow cleavage and bedding. This second type of fold axis is by far the most prevalent (Pl. 1).

West and north of Sugarloaf Mountain, the axes of small folds tend to plunge under the mountain (Pl. 1). This led some previous workers to believe the mountain was synclinal (Cloos and Hietanen, 1941, p. 96; Stose and Stose, 1946, p. 121). However, more positive evidence indicates the structure is anticlinal. Also the lineation which expresses the axes of the more important folds in the phyllite almost invariably plunges away from the mountain (Pl. 1).

These larger fold axes show a variation in strike north of the north nose of the Sugarloaf

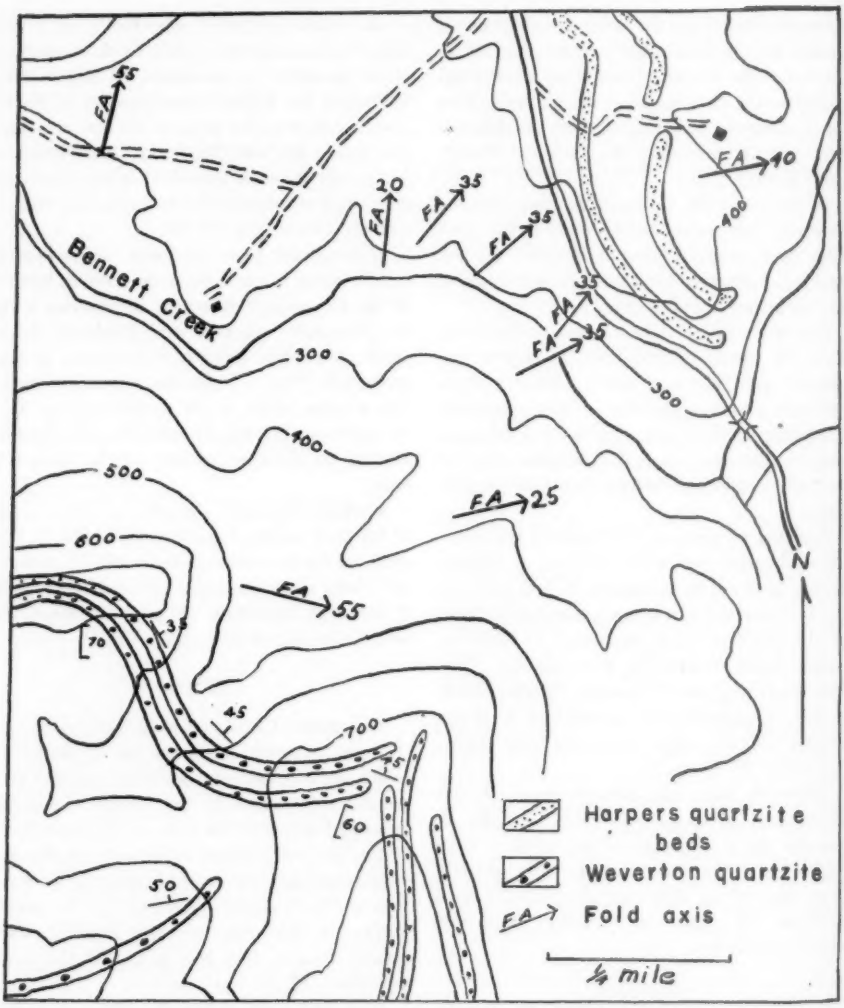


FIGURE 6.—DETAIL MAP OF NORTH NOSE OF SUGARLOAF ANTICLINE
Showing changing orientation of fold axes in Harpers phyllite.

anticline. On the east and west sides of the mountain, these axes strike parallel to the axis of the large anticline. However (Fig. 6), north of the mountain their orientation becomes more easterly.

The change in the orientation of fold axes is probably the result of overturning of the large anticline (Fig. 7). Since the axial plane dips to the east, the fold axes trend more easterly than normal near the nose of the anticline. Furthermore, the center portions of the fold are moved farther forward during folding than the flanks (Fig. 9). The shear produced by this differential movement between the large fold and the phyllite caused a bending of the fold axes in the phyllite to a more east-west orientation. The orientation of the fold axes in the phyllite which are not opposite the nose of the large

fold but near the limbs was not changed because there was no differential movement; the whole limb moved westward as a unit.

Faults

Minor displacements.—Several east-west displacements of a few hundred feet occur in the quartzite beds both on the mountain and in the Harpers phyllite. A mile and a quarter north of the south end of the west ridge, two beds of Weverton quartzite are displaced 300 feet. The northern beds moved west. The largest displacement is on the east limb of the plunging syncline of quartzite in the Harpers phyllite a mile southwest of Urbana (Pl. 1). Here the northern bed has been moved a quarter of a mile to the northwest. Only the horizontal component of the displacement could be measured in these faults.

Major faults.—A possible reverse fault has been mapped (Pl. 1) as extending from the Potomac River to the northern boundary of the map at an azimuth of about 30°. It follows the western contact of the Harpers phyllite.

Two lines of evidence point to the possible existence of this fault. First, a mile west of Flint Hill along the west bank of the Monocacy River, the Frederick limestone is in contact with the Harpers phyllite. Since these two formations are normally separated by the Antietam quartzite, the contact must be a fault, the eastern block having moved up in relation to the western block. An eastward dip to the fault is assumed since all other structures in the area dip to the east. If the assumption is correct, the fault is a reverse fault.

The second line of evidence involves the extension of the trace of this fault north and south, along the contact between the Harpers phyllite and Antietam quartzite which lies directly to the north and south (Pl. 1). This contact is nowhere exposed, but there is some evidence of faulting. The lineation which lies in the cleavage of the Harpers phyllite strikes normally about 105°. The lineation becomes markedly more southerly of this just east of the Harpers-Antietam contact south of the Harpers-Frederick limestone fault contact. Also, a less pronounced alteration is noted in the direction of flow-cleavage strike from the normal 15° or 20° to more westerly directions near the contact. A similar change in the orientation of foliation and lineation in response to thrust faulting in Duchess County, New York, has been described by Balk (1936, p. 74).

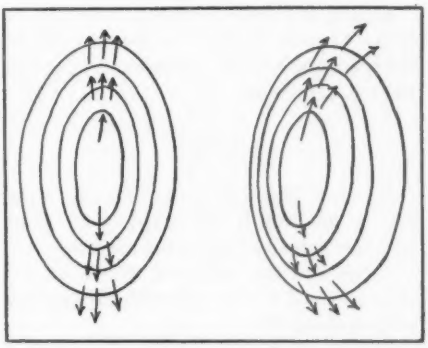


FIGURE 7.—SCHEMATIC MAP OF ANTICLINAL DOME SHOWING ORIENTATION OF FOLD AXES BEFORE (LEFT) AND AFTER OVERTURNING (RIGHT)

Jonas and Stose (1938) mapped this line as part of the "Martic Overthrust." The writer believes that, although there is evidence of faulting and an overthrust is possible, it is more likely the faulting is only local and displacement has not been of overthrust proportions. The Harpers phyllite on either side of the possible fault line has very similar lithologic characteristics and grade of metamorphism. When Sugarloaf Mountain was believed to be a syncline, it seemed logical that a thrust of large proportions separated it from the also synclinal Frederick Valley. All evidence indicates, however, that Sugarloaf Mountain is an anticline, and it is possible, but no longer necessary, that an overthrust separates the mountain from the valley.

Joints

Although a special study of the numerous joints was not undertaken, some prominent trends in the Weverton quartzite are apparent. Cross jointing (*ac*) is the most common type. These joints cut the cleavage at a high angle having a generally southeast strike and steep northerly dip. In some places the joints are closely spaced giving the appearance of a

fracture zone. This jointing is parallel to the small east-west faults common in the area.

Two other joint systems are readily observed in the Weverton. Both parallel the usual 20-degree strike of the bedding. One dips steeply to the west; the other has a gentle eastern dip.

Jointing is less pronounced and less regular in the Harpers phyllite. No trends could be determined without an exhaustive study, which was not undertaken.

Folding

Sugarloaf Mountain anticline.—The major fold in the area is the Sugarloaf Mountain anticline. It is an asymmetrical anticlinal dome overturned to the northwest. The dome is 5 miles long and 2 miles wide with the long axis striking about 30°.

Earlier workers believed that the structure was a monocline (Keyes, 1891) or two compressed synclines (Stose and Stose, 1946), because almost everywhere on both ridges of the mountain the trend of bedding is parallel. Detailed mapping of the quartzite ridges revealed a change in strike and dip and a connection between the two ridges of the mountain at the south as well as north end.

The topography of the mountain is misleading since the two ridges are joined topographically at the north end but not at the south. However, quartzite beds which form a bench below the main peak on the east ridge may be traced southwest to a point where they coincide with those of the west ridge to form the south nose of a plunging anticline (Pl. 1).

The structure is also joined at the north. In the valley between the eastern and northern ridges of the mountain, two resistant quartzite beds express the structure as two plunging anticlines on either side of a plunging syncline. More accurately it may be thought of as a single plunging anticline with a synclinal flexure at the crest.

The dome plunges about 30° to 35° at both the north and south noses. The bedding at the nost of the western flexure of the north end of the anticline strikes northwest and dips 35° NE. At a point where the nose is apparently exposed at the south end of the structure, the beds strike east-west and dip 35° S.

The dip of the bedding on the eastern limb is generally not so steep as that on the overturned western limb. This accounts to some extent for the greater width of the eastern ridge.

Sugarloaf Peak and the peak north of it consist of secondary folds. They are apparently flexures or drag folds formed in a single bed within the east limb of the larger fold (Fig. 8). The quartzite bed which forms the main peak has been contorted into complex minor folds, plunging steeply east. The beds on either side of the contorted bed are not folded. On the next peak to the north a simpler flexure occurs. It is a single tight fold plunging to he east with its axis at a high angle to that of the main fold. Here again the adjacent bed is not affected by the folding but maintains a constant trend opposite the folded zone. This tight minor folding thickens the section and accounts for the existence of the peaks.

Mechanism of folding.—A cross section constructed to scale from a measured section is shown in Figure 10.

The Sugarloaf Mountain anticline is probably a shear fold which has been modified by flowage in the manner described by Cloos (1947, p. 902). The shearing and attendant flowage took place parallel to the cleavage. Since bedding and cleavage are frequently parallel in the west limb of the fold, shearing tended to thin the quartzite beds on that limb. Similarly, shearing thickened the beds on the east limb, since here the cleavage is always steeper than the bedding.

A further result of shearing is seen on the west limb of the anticline, where it is well exposed on the east side of the valley cut by the Bear Branch at the northwest side of the mountain. A small fold, 15 feet wide and 10 feet high, is exposed (Pl. 5, fig. 2). The axis of the fold strikes 30° and plunges gently northeast. The axial plane dips 65° E. and thus parallels the bedding and cleavage in the unfolded quartzite exposures on either side of the small fold.

The fold was produced by movement between adjacent quartzite beds. The eastern, or inner, beds of the larger anticline must have moved upward in relation to the western, or outer, beds of the anticline. Such movement is opposite that produced by flexure folding in which the

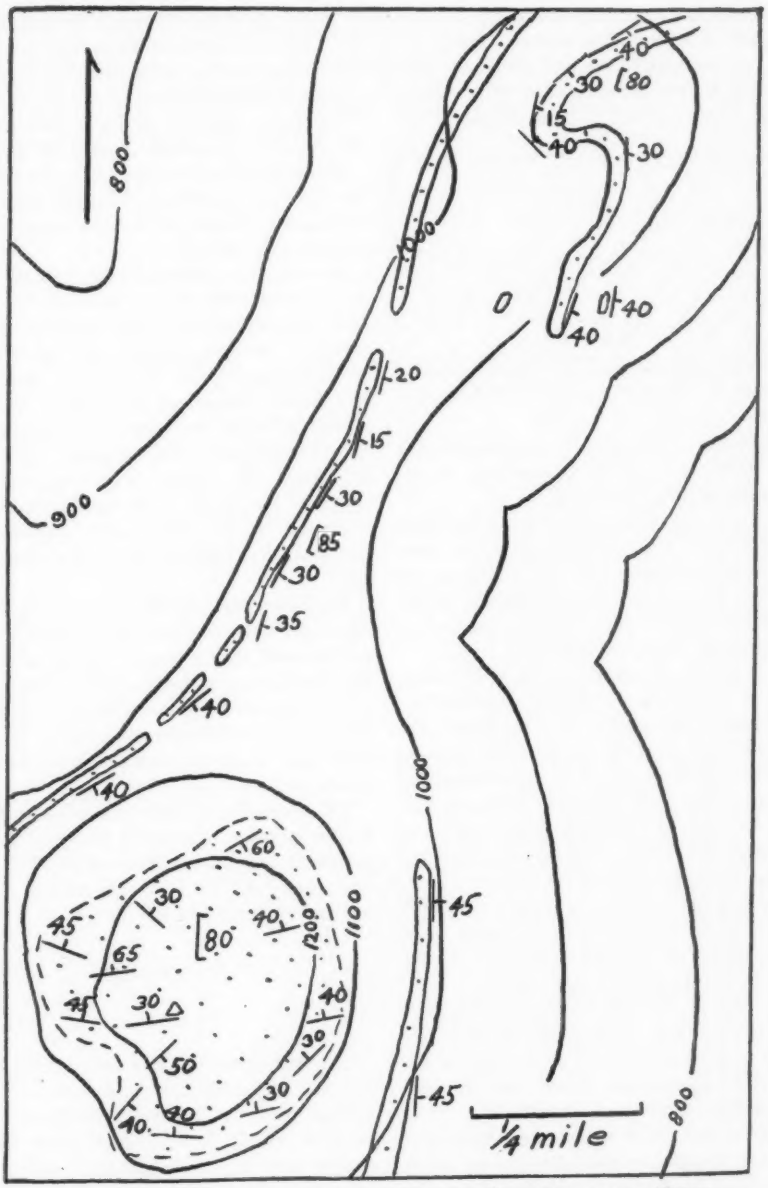


FIGURE 8.—DETAIL MAP OF SUGARLOAF PEAK AND NEXT PEAK TO THE NORTH
Showing erratic orientation of bedding in the contorted quartzite beds comprising these peaks.

outer beds of an anticline move upward in relation to the inner beds (Hills, 1943, p. 91). By itself, this minor fold would seem to indicate that the larger fold is a flexural syncline. How-

ever, so much other evidence indicates that the large fold is an anticline that this small fold must be interpreted as the result of drag produced during the formation of the large anticlinal shear fold. The inner portion of the

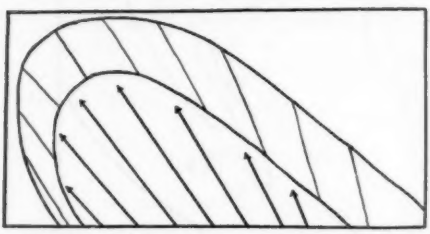


FIGURE 9.—MECHANISM OF SHEAR FOLDING

anticline was thrust farther upward than the outer portion, as would be expected in the formation of such a fold (Fig. 9).

In many ways this small fold is a duplicate of the larger one. It is overturned to the west; the cleavage is steeper than the bedding on the east limb and parallel to the bedding on the west limb; the west limb dips more steeply than the east limb and is overturned.

Folding in the Antietam quartzite.—The western belt of Antietam quartzite is flanked on either side by the younger Frederick limestone. The belt must therefore be anticlinal. Further indication of this is found in measurements of bedding where a cross section of the fold is exposed 1 mile east of Buckeystown. Here the Monocacy River swings abruptly east cutting a valley through the Antietam ridge.

The fold is about half a mile wide and 9 miles long extending from the Potomac River northeastward to a point just north of the northern boundary of the map (Pl. 1).

The eastern belt of Antietam quartzite is discontinuous and parallels the western belt half a mile to the east. It is flanked by the Frederick limestone on the west and the Harpers phyllite on the east. Its structure is less clear than that of the western belt. Possibly it is also an anticline, in which case the Antietam would be in fault contact with the older Harpers phyllite on the east. The contact between these two rocks is nowhere exposed so their relationship remains obscure. However, the bedding which can be observed in exposures on the northeast side of the Monocacy River at Greenfield Mills indicates that the belt at this point consists of a series of tight folds. As the contact with the Harpers phyllite east of this outcrop is not exposed, it cannot be determined whether the Antietam dips above or below the Harpers.

This contact has been mapped as the trace of the "Martic Overthrust" by Jonas and Stose (1938). The present writer has mapped it as a possible fault contact.

Folding in the Harpers phyllite.—In the Harpers, the phyllitic layers have been deformed as incompetent strata between the more competent quartzite beds. Although the folding of the quartzite beds within the Harpers phyllite is not readily determined, it is undoubtedly of relatively broad proportions. A syncline of quartzite plunges to the northwest a mile west of Urbana. The outcrop of the fold is 2 miles long and a mile wide. Its synclinal nature is indicated by fold axes within the adjacent phyllitic beds which plunge beneath the quartzite.

The incompetent phyllite has probably flowed into tight folds. Although the phyllite is fairly well revealed, the exposures are small, and these folds are never completely exposed. Their presence is indicated by the different attitude of bedding in near-by outcrops, as in the phyllitic exposures along the north side of Bennett Creek for 5 miles east of Park Mills.

The orientation of these folds is best reflected in the attitude of lineation produced by the intersection of cleavage and bedding. This lineation parallels the fold axes and has been mapped wherever possible (Pl. 1).

PETROFABRICS

Introduction

The structural problem has been largely attacked megascopically, and it was not felt that an exhaustive petrofabric study was necessary. In the hope of gaining some further insight into the mechanisms of folding and deformation in the Sugarloaf anticline a limited petrofabric study was made of the Weverton quartzite. Twelve specimens were taken from various locations around the anticline (Fig. 12), and one specimen was selected from the quartz-

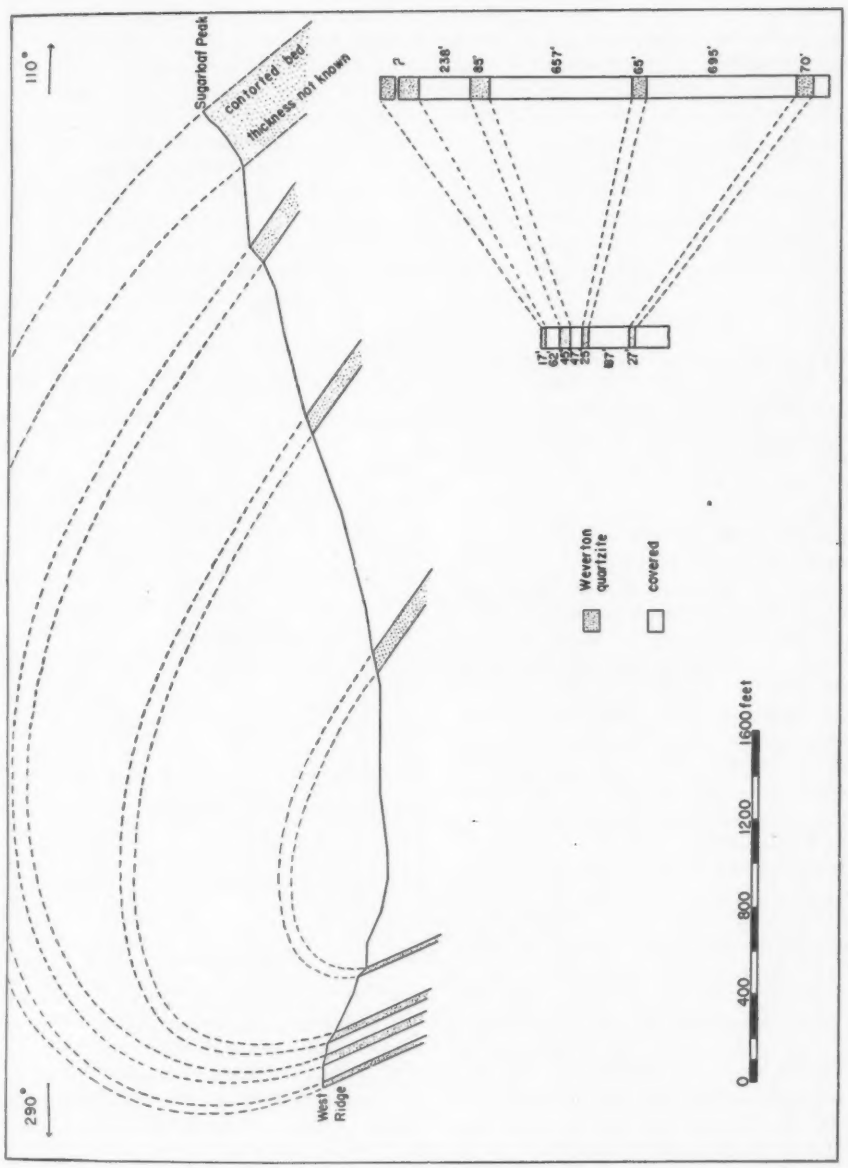


FIGURE 10.—SECTION THROUGH THE SUGARLOAF ANTICLINE CONSTRUCTED FROM TRAVERSE

ite in the Harpers phyllite for comparison. Thirty-five petrofabric diagrams are presented here. They consist of the projections of poles of mica cleavages and planes of liquid inclusions and the *c* axes of quartz in the lower hemisphere of an equal-area net. The geographic orientation of the diagrams is indicated by the strike and dip symbol on the diagram. Dips less than 90° indicate a plane dipping toward the reader, and dips greater than 90° are overturned and indicate a plane dipping away from the reader.

Mica Orientation

All mica diagrams show the orientation of mica flakes in an *ac* girdle of varying degrees of completeness. As has been demonstrated by E. Cloos (Cloos and Hietanen, 1941, p. 41), the extent of girdle is related to the height of maxima in the girdle.

With two exceptions, there is more than one maximum within the girdle indicating the presence of more than one *S* plane. Except in diagram D1, where a few mica flakes have accumulated in a fracture producing a small maximum in the girdle, all these planes are considered flow cleavage planes—S_2. This seems valid since no fracture cleavage was noted megascopically in these specimens, and in any case would probably not be expressed by mica orientation (Cloos, 1941, p. 40). Since no mica maxima correspond with the poles of bedding planes measured megascopically, evidently no mica flakes lie in the bedding.

In diagram B1, no mica maximum occurs at the pole of the megascopic cleavage. However, this megascopic cleavage is revealed microscopically by an elongation of quartz grains. In a few other cases the megascopic cleavage is not accompanied by an alignment of mica, but its orientation has been determined on the basis of the elongation of quartz grains visible megascopically. There is indication that the *c* axes of quartz lie in this megascopic cleavage plane which also contains the axes of elongation of quartz grains. Since quartz is more sensitive to deformation than mica (Sahama, 1936) and may therefore represent the latest deformation, the megascopic cleavage appears to be the latest.

The youngest *S* plane, that which corresponds to megascopic cleavage, is designated Meg. Cl. on the diagrams although it is also an S_2 plane. The oldest flow cleavage plane is called S_2, the intermediate one $S_{2'}$.

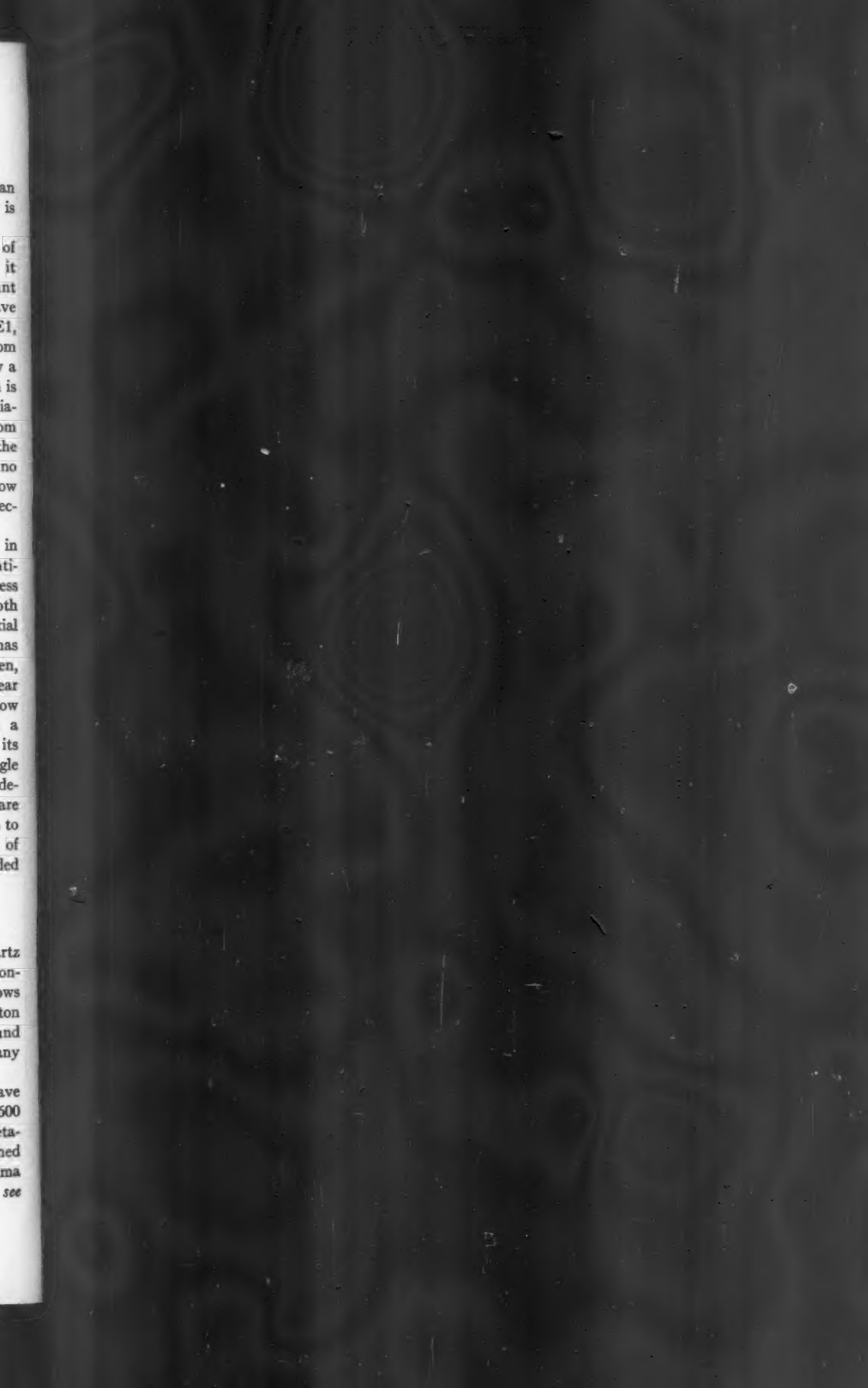
Knowing which cleavage is the younger of those indicated by the petrofabric diagram, it is possible to determine the minimum amount and direction of rotation these *S* planes have undergone (Table 3). Diagrams B1, C1, D1, E1, and F1 were made from specimens taken from the east limb of the anticline. They all show a counter-clockwise rotation when the diagram is viewed from the south toward the north. Diagrams G1, H1, I1, and J1 were made from specimens taken from the west limb of the anticline. Two of these (H1 and I1) show no rotation of *S* planes. Diagrams G1 and J1 show a 9- and 50-degree clockwise rotation respectively.

Thus, apparently the rotation has been in opposite directions of either limb of the anticline. Although the evidence of rotation is less clear on the western limb, rotation on both limbs seems to have been toward the axial plane of the fold. This same relationship has been noted by Hietanen (Cloos and Hietanen, 1941, p. 165) in the Wissahickon schist near Bryn Mawr, Pennsylvania. Evidently the flow cleavage has rotated during folding from a position at a high angle to the bedding to its present orientation parallel to or at a low angle to the bedding. Degrees of rotation as determined from the petrofabric diagrams are subject to too many errors and irregularities to be reliable, but the consistency of direction of rotation on the two limbs of the fold is regarded as significant.

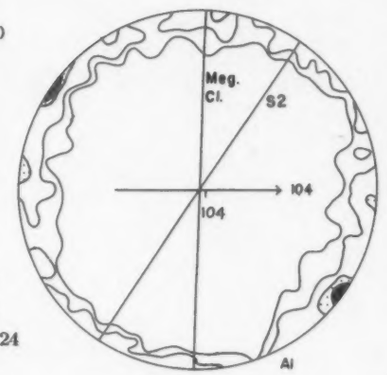
Quartz Orientation

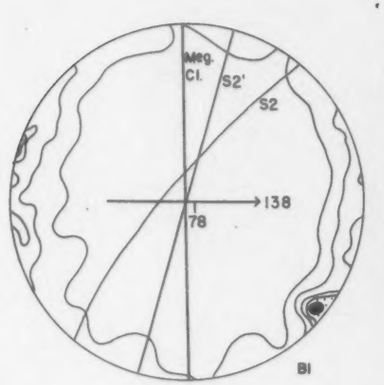
Strong maxima are lacking in the quartz diagrams. Only one diagram (Pl. 6, B3) contains a maxima greater than 4 per cent. Fellows (1943, p. 1427) made two diagrams of Weverton quartzite from Sugarloaf Mountain and found the lowest degree of quartz orientation of any tectonite he studied.

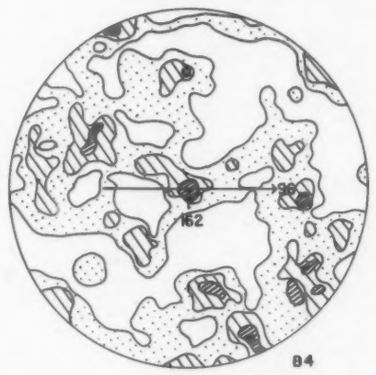
Thirteen partial diagrams of quartz have been prepared for which the orientation of 2600 quartz grains was determined. In the interpretation of the diagrams, the maxima were matched with the positions of theoretical quartz maxima as designated by Fairbairn (1942, Fig. 123; *see*

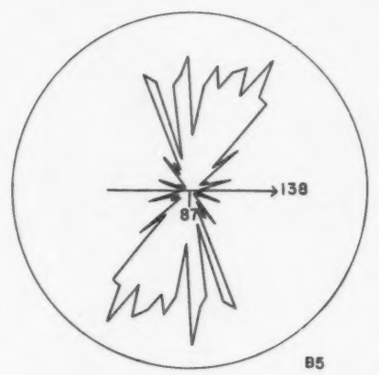


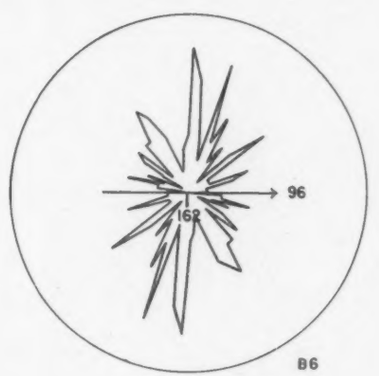
Specimen	Diagram	Mineral	No. of grains	Contours (per cent)
25	A1	mica	100	2, 4, 8, 12
25	A2	large quartz	200	1, 2, 3
25	A3	small quartz	200	1, 2, 3, 4
28	B1	mica	100	1, 4, 8, 10, 12
28	B2	large quartz	200	1, 2, 3, 4
28	B3	small quartz	200	1, 2, 5, 6
28	B4	large quartz	200	1, 2, 3, 4
28	B5	elongation of quartz	200	two dimensional
28	B6	elongation of quartz	200	two dimensional
28	B7	planes of liquid inclusions	50	4, 10, 14, 18, 22, 24
33	C1	mica	100	1, 5, 7, 11, 13
33	C2	large quartz	200	1, 2, 3
33	C3	small quartz	200	1, 2, 3, 4
33	C4	planes of liquid inclusions	50	2, 6, 12
68	D1	mica	100	2, 5, 8, 11, 14
68	D2	large quartz	200	1, 2, 3, 4
68	D3	planes of liquid inclusions	50	4, 6, 8, 10, 12
27	E1	mica	100	1, 4, 8, 12

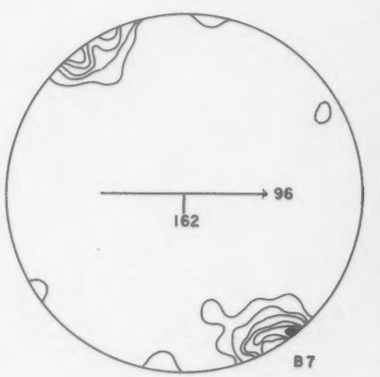


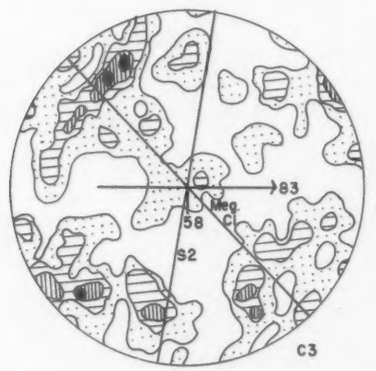


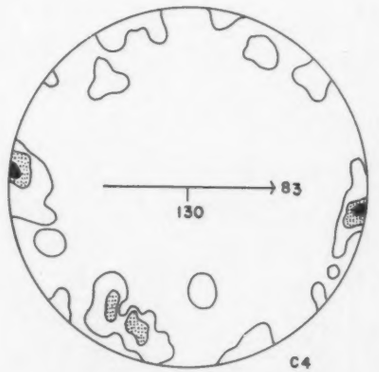


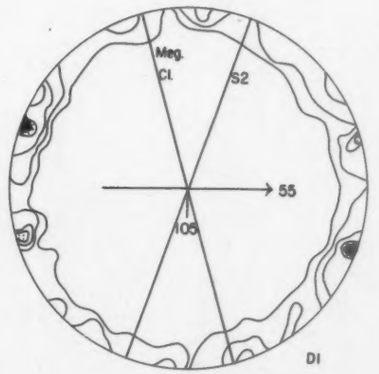




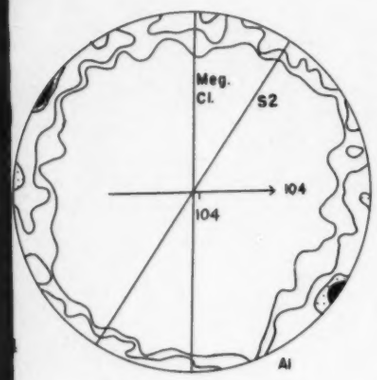


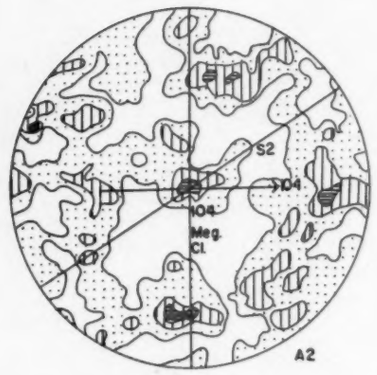


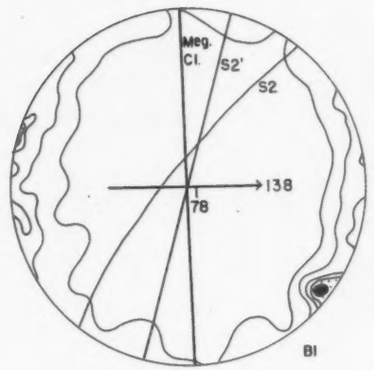


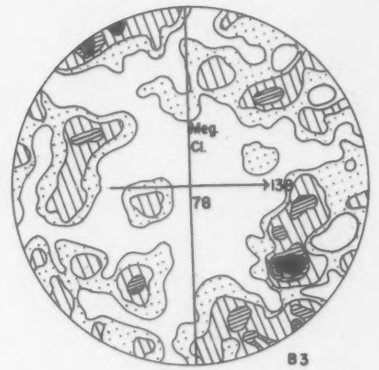


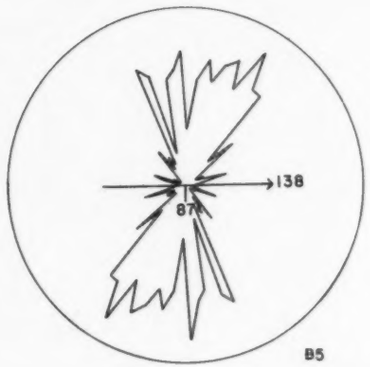
PETROFABRIC DIAGRAMS

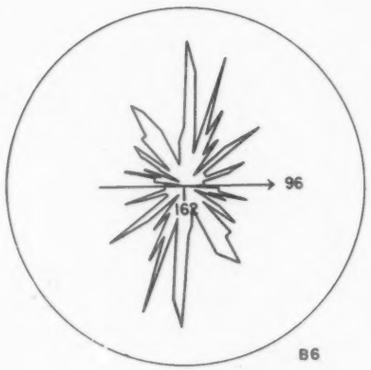


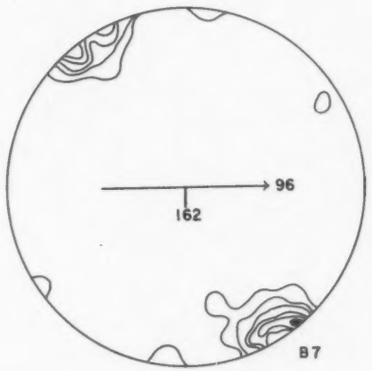


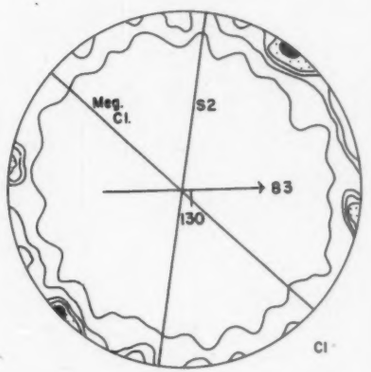


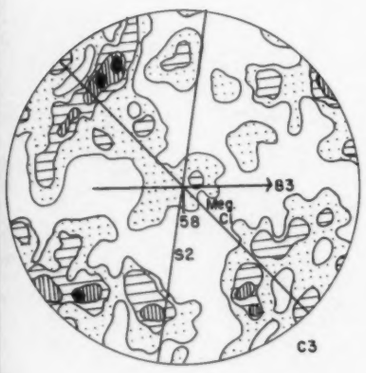


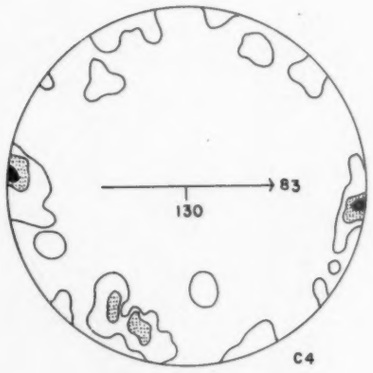


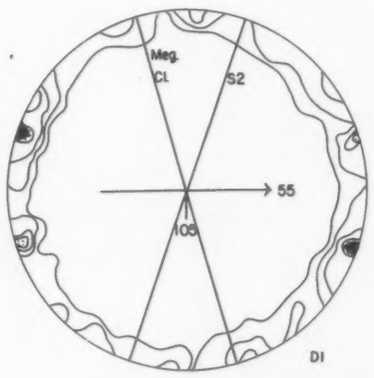


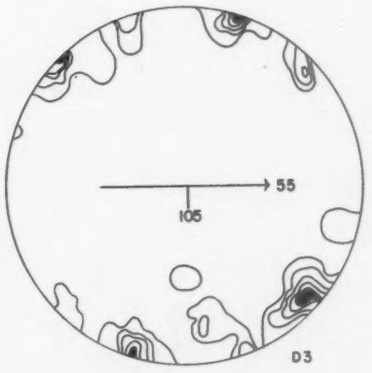


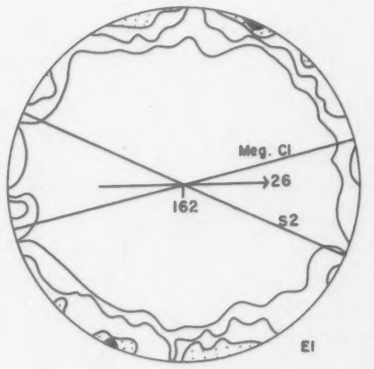












PETROFABRIC DIAGRAMS

TABLE 3.—DESCRIPTION OF MICA DIAGRAMS

Specimen number	Diagram number	Width of girdles on periphery in degrees	Maxima per cent	Degrees and direction of *S* plane rotation	Remarks
25	A1	1 4 360 315	12, 8	77 counterclockwise	A B-tectonite with rotation around B.
28	B1	1 4 350 170	12, 10	40 counterclockwise	Megascopic celavage has no mica maxima normal to it, but is expressed microscopically by a quartz elongation.
33	C1	1 5 360 190	13, 11, 11	52 counterclockwise	A B-tectonite with rotation around B′.
68	D1	1 5 350 190	14, 8	37 counterclockwise	Also an 8% maximum at pole of fracture. A B-tectonite with rotation around B′.
27	E1	1 4 350 210	12, 8	39 counterclockwise	A B-tectonite with rotation around B′.
53	F1	1 4 295 130	15, 15, 12	26 counterclockwise	A B-tectonite with rotation around B.
66	G1	1 5 260 140	18, 16	9 clockwise	Megascopic cleavage strong. Therefore maximum is large, girdle is small, and rotation is not large.
42	H1	1 5 350 180	16, 13	no rotation	13% maxima are symmetrically arranged on either side of the main 16% maxima. Therefore no definite rotation determinable.
41	I1	1 6 347 150	18	no rotation	The one significant maxima allows no determination of the direction of rotation.
41	J1	1 4 300 180	12, 12 12, 12	52 clockwise	Four stages of *S* plane rotation produce a B-tectonite.

Fig. 11). The *b* axis is normal to the page with the *a* axis horizontal and the *c* axis vertical. The megascopic cleavage in the quartzite is considered to lie in the *ab* plane. Thus the center, or *B* axis, of the petrofabric diagram was placed over the center of the key diagram and the megascopic cleavage plane rotated to the horizontal. Small adjustment was usually made to bring the actual maxima into coincidence with the theoretical positions, but no maxima which fell more than 10° from a theoretical maximum were considered correlative.

Most of the diagrams show an incomplete and irregular girdle about 20°–30° from the periphery. This cleft girdle pattern is described by Fairbairn (1939, p. 1487) as typical of B-tectonites. The frequency of standard maxima is indicated in Table 4. Aside from maxima which correspond to no standard, position VII

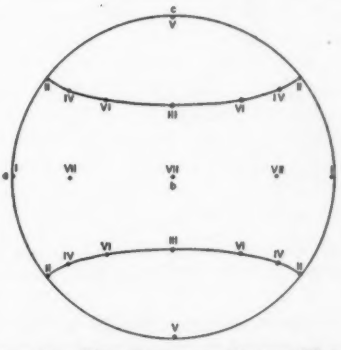


FIGURE 11.—DESIGNATION OF QUARTZ MAXIMA (After Fairbairn, 1942).

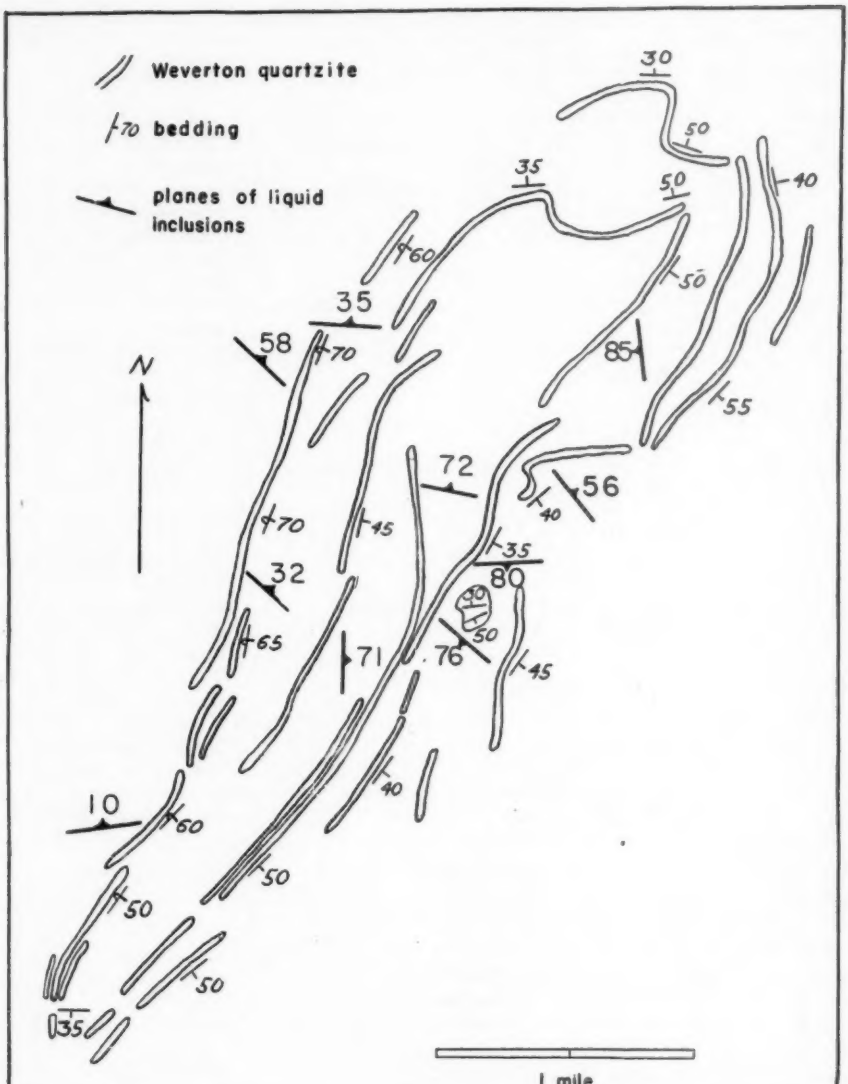


FIGURE 12.—SKETCH MAP OF SUGARLOAF MOUNTAIN ANTICLINE SHOWING LOCATION OF SPECIMENS

is the most common. Positions I, VII, and VIII, which all lie in the *ab* plane, account for 13 maxima while only 9 maxima fall in all other standard positions. Thus the *c* axis of quartz is apparently most commonly oriented in the *ab* plane which is also the megascopic cleavage plane. Since quartz expresses the latest deformation movements, this provides some indication that megascopic cleavage is the latest and tends to support the conclusion reached as to the direction of *S*-plane rotation.

It is realized that a deformation does not

necessarily produce orientation of the *c* axis of quartz within an *S* plane which is also a product of that deformation. In Fairbairn's (1942, p. 69) order of frequency of occurrence of standard quartz maxima, maximum VII was among those occurring least frequently. However, Cloos and Hietanen (1941) found position V, also one of those to be expected least often, to be among the most common in the quartz fabric of the Wissahickon schist and related rocks in Maryland and Pennsylvania.

The conclusion that quartz axes have been oriented in the *ab* plane by the last deformational movements may possibly be based on too few data. Several diagrams were prepared and discarded because the position of tectonic axes, *a*, *b*, and *c*, could not be determined, thus no identification of maxima could be obtained. The fact that so few maxima above 4 per cent occur may diminish the validity of the conclusion somewhat.

Diagrams B2 and B4 (Pl. 6) were made from sections of the same specimen cut normal to each other. The 4 per cent maximum near B in diagram B2 also appears in B4, indicating that the two diagrams are rotatable to that extent. However, no other maximum of 4 per cent or higher occurs in diagram B4.

TABLE 4.—POSITIONS OF QUARTZ MAXIMA

Diagram number	Specimen number	Maxima according to Fairbairn (1942) (percentage)									Remarks
		I a	II	III	IV	V c	VI	VII	VIII b	N.C.*	
A2	25							4	4	4	Good cleft girdle lineation in *b*
A3	25	4									Small quartz, good girdle
B2	28				4				4		Poor cleft girdle
B3	28						6			5	Small quartz, largest maxima
B4	28								4		From section cut normal to B2, common 4% maximum at VIII
C2	33		no maxima above				3	3		3	Poor orientation
C3	33						4	4		4	Small quartz, fair cleft girdle
D2	68		4			4		4			Fair girdle
F2	53				3			3		3	Poor orientation, no maxima above 3%
G2	66			4							Orientation not strong
H2	42				4					4	Fair cleft girdle
I2	36							4		4	Fair girdle
J2	41	4						4			Good girdle
Total.....		2	1	1	3	1	3	7	3	7	

* No correlation.

Quartz Elongation

The elongation of original quartz grains, studied microscopically, are illustrated in diagrams B5 and B6, made from sections cut normal to each other from the same specimen. A measurement of 200 grains was made in each slide. The elongation lies generally in the megascopic cleavage plane (B5). A comparison of diagram B5 with the mica diagram from the same slide, B1, shows a remarkable similarity between the poles of the direction of elongation of quartz and mica maxima. Evidently the quartz tends to become elongated in the S_2 planes. The total effect of this elongation is to express the megascopic cleavage.

The ratio of elongation is 1:1.7 in both sections, indicating that the quartz grains are disc-shaped. A comparison of diagrams B5 and B6 with the corresponding large quartz diagrams B2 and B4 from the same sections shows that the 4 per cent maxima in B which is

present in both quartz diagrams lies close to the elongation plane. The other 4 per cent maxima, present in diagram B2 only, does not lie in this plane.

There is some indication in these diagrams, then, that the *c* axes of quartz tend to lie in the cleavage plane. This supplements the same conclusion drawn from the prevalence of maxima of all quartz diagrams lying in position I, VII, and VIII.

Planes of Liquid Inclusions

During the petrofabric study, the recent paper by O. F. Tuttle (1949) appeared. Since the Washington, D. C., area studied by Tuttle lies about 30 miles southeast of the Sugarloaf Mountain area, it was felt that a study of the planes of liquid inclusions in the Weverton quartzite on Sugarloaf Mountain (Pl. 4, fig. 1) might prove of value for comparison purposes. Many of the thin sections used in the mica and quartz petrofabric studies were reinvestigated. As these sections were originally prepared for the petrofabric study, many were not cut properly to reveal the orientation of these planes. From over 30 slides available, only 10 satisfactory diagrams could be prepared. These planes show a high degree of orientation within a single specimen, so that no large number of measurements was necessary. Fifty measurements per slide were sufficient.

According to Tuttle, the planes of inclusions resulted from the fracturing of quartz grains and the later filling of the fractures with solutions. Through a process of solution and deposition of silica, the solution-filled fractures become changed to numerous liquid inclusions lying in the original fracture plane. These planes bear no constant relationship to the crystallographic direction in quartz, indicating an isotropic behavior of quartz during the fracturing.

In the Washington, D. C., area, Tuttle found a remarkably uniform orientation of the planes of liquid inclusions, striking northwest and dipping steeply northeast. No megascopic structures were genetically related to these planes; their orientation was invariably different.

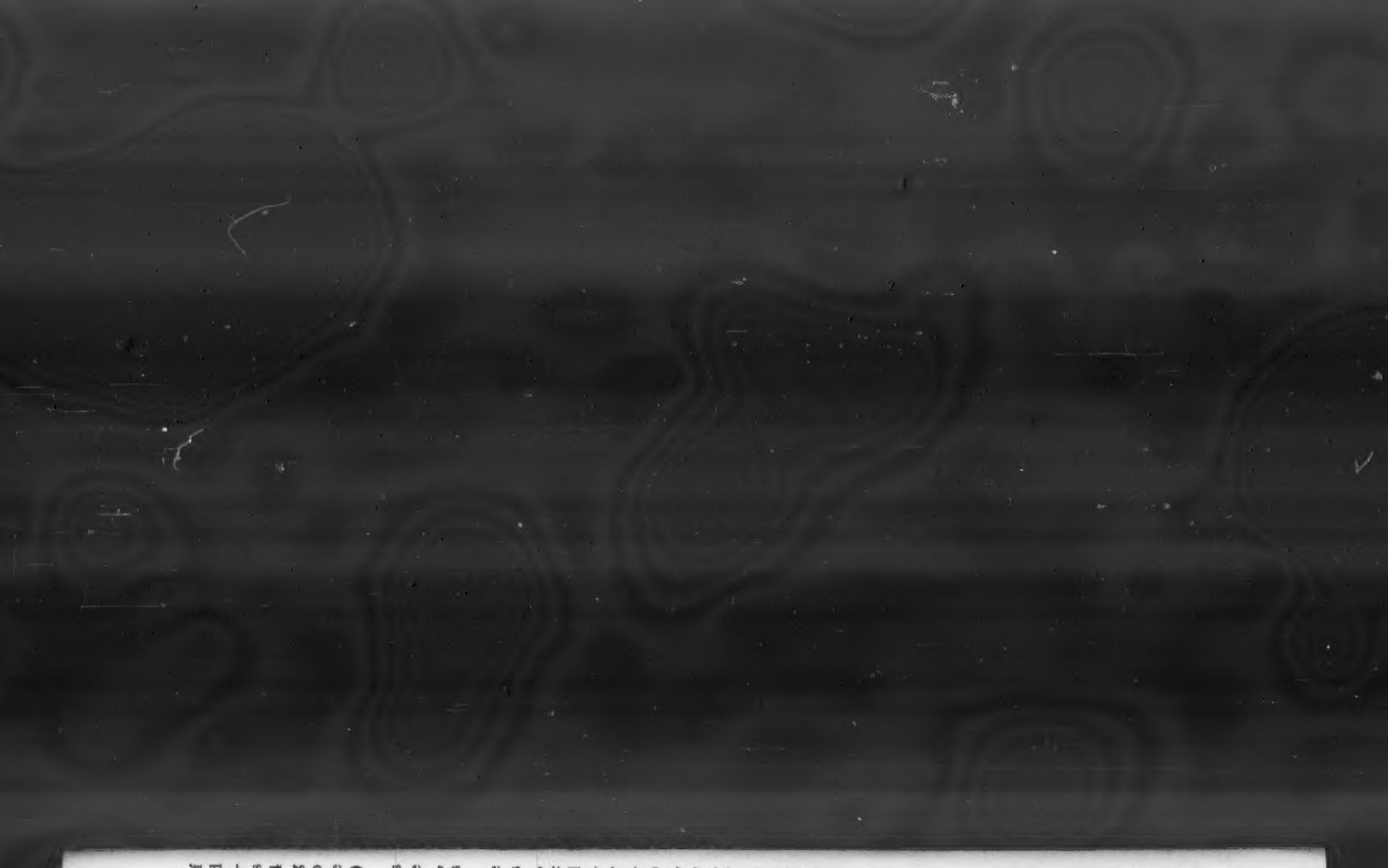
In the Sugarloaf area, no crystallographic control of the planes of liquid inclusion was found. This is illustrated by a comparison of diagrams B7, C4, D3 (Pl. 6) and G3, H3, and J3 (Pl. 7) with the corresponding quartz diagrams. Only in diagrams B7 and J3 do the poles of planes of liquid inclusions coincide with 4 per cent quartz maxima in the corresponding quartz diagrams B4 and J2, respectively. Also no similarity between these planes and the theoretical quartz maxima of Fairbairn (Fig. 11) was noted.

No constant relationship between the poles of these planes and the poles of mica cleavage seems to exist. Only in diagrams C4 and C1, made from the same specimen, do the two poles coincide.

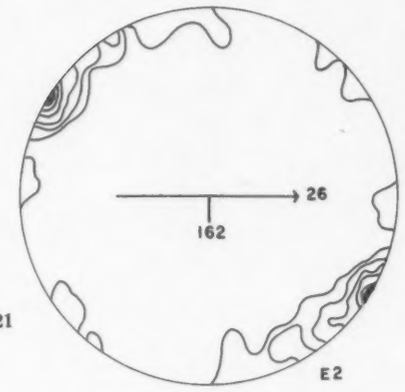
Tuttle found no relationship between the orientation of the planes of liquid inclusions and megascopic structure in the Washington, D. C., area. Figure 13 shows the geographic position and orientation of the planes of liquid inclusions in the quartzites of Sugarloaf Mountain. The planes have been reoriented into their geographic attitude from the petrofabric diagrams. The plane reoriented was that normal to the chief maxima in each diagram. All but a few of the planes strike 120° to 130°. The dip is variable, but a northeasterly one predominates. The same thing is shown in diagram N in which the poles of the planes of liquid inclusions have been plotted in the horizontal plane with north at the top. The band of points extending from northeast to southwest reveals this consistency of strike. The light concentration of points in the southwest quadrant illustrates the more usual northeast dip.

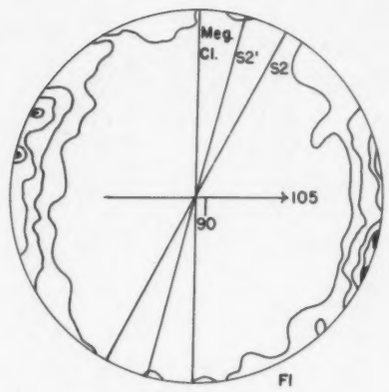
The only megascopic structural feature which this tends to parallel is the dominant *ac* cross jointing, which has an average strike of about 120°, and a northerly dip.

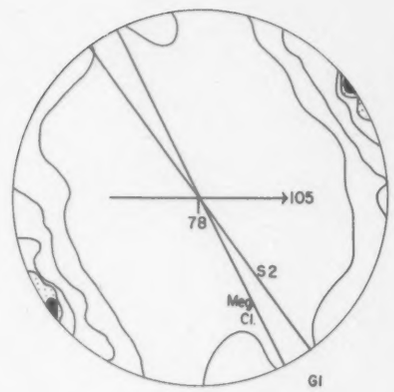
The results, though not as conclusive as would be possible with fuller data, do indicate a probable relationship between the megascopic structure and these microscopic planes. Tuttle points out that no *ac* joints occur in the Washington, D. C., area so that their relationship to the planes of liquid inclusions could not be determined. He believes, however, that if there is a relationship, possibly megascopic jointing is the shallow-zone manifestation of the movement which produced planes of liquid inclusions at greater depths. The Sugarloaf Mountain area

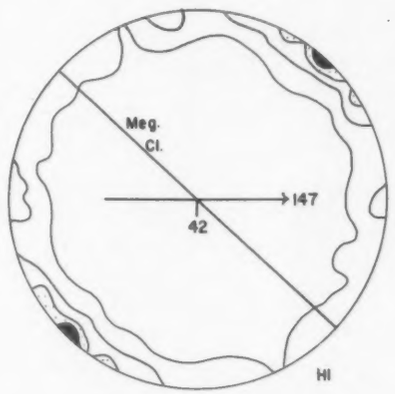


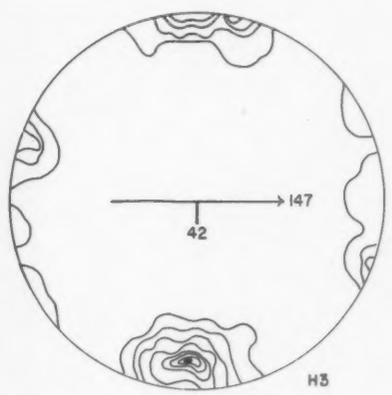
Specimen	Diagram	Mineral	No. of grains	Contours (per cent)
27	E2	planes of liquid inclusions	50	4, 8, 12, 16, 20, 24
53	F1	mica	200	1, 4, 6, 10, 14
53	F2	large quartz	200	1, 2, 3
66	G1	mica	100	1, 4, 8, 14, 18
66	G2	large quartz	200	1, 2, 3, 4
66	G3	planes of liquid inclusions	50	2, 4, 12, 16, 20
42	H1	mica	100	2, 5, 13, 16
42	H2	large quartz	200	1, 2, 3, 4
43	H3	planes of liquid inclusions	50	2, 5, 8, 11, 14, 17, 19, 21
36	I1	mica	100	1, 6, 10, 14, 18
36	I2	large quartz	200	1, 2, 3, 4
41	J1	mica	100	1, 4, 8, 12
41	J2	large quartz	200	1, 2, 3, 4
41	J3	planes of liquid inclusions	50	4, 8, 14, 16, 20, 24
63	K1	planes of liquid inclusions	50	2, 6, 10, 14, 22
65	L1	planes of liquid inclusions	50	2, 8, 14, 20, 30, 32
62	M1	planes of liquid inclusions	50	2, 8, 14, 18, 22, 28
	N	compilation diagram of planes of liquid inclusions		

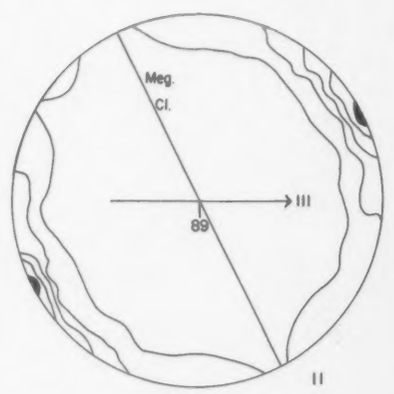


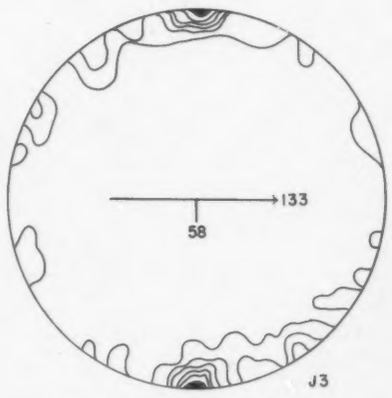


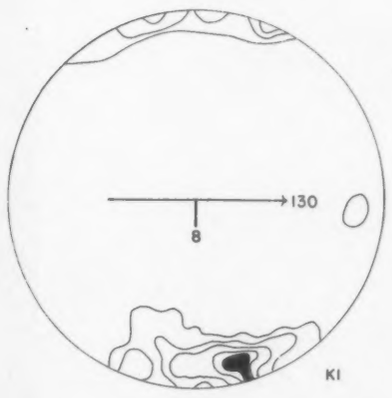


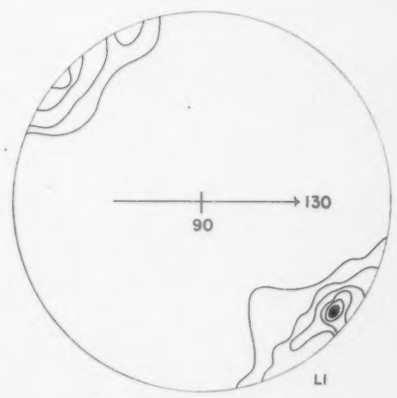


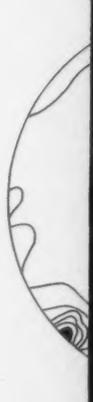




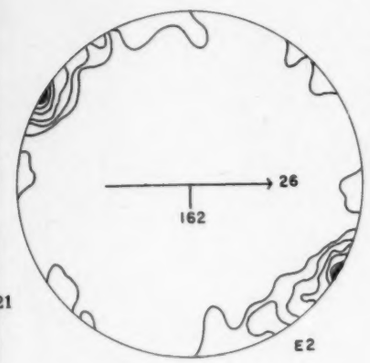




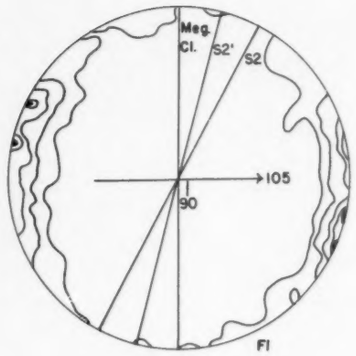




PETROFABRIC DIAGRAMS

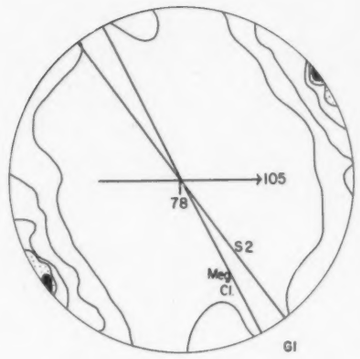


E2



F1

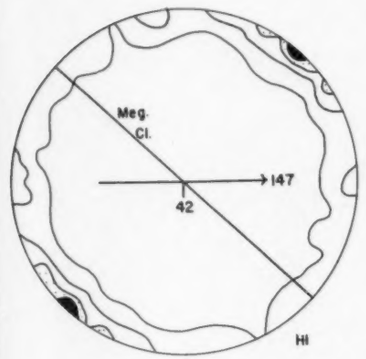
F2



G1

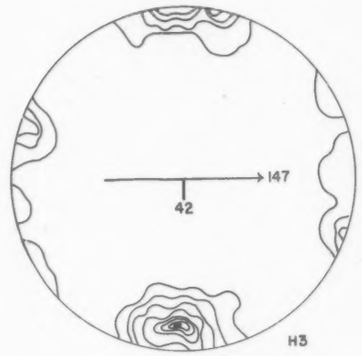
G2

G3

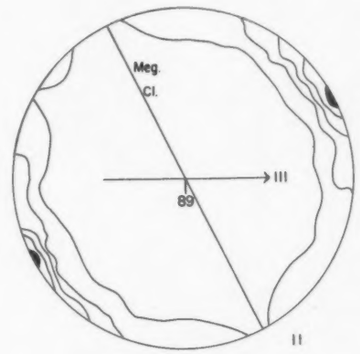


H1

H2

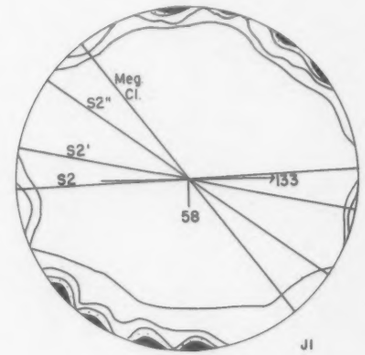


H3



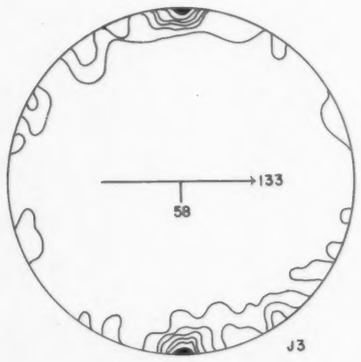
I1

I2

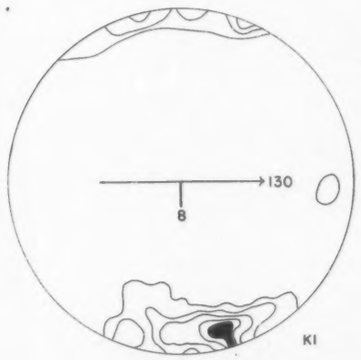


J1

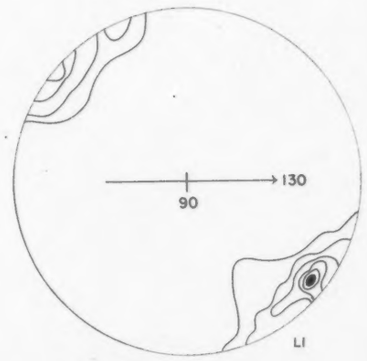
J2



J3



K1



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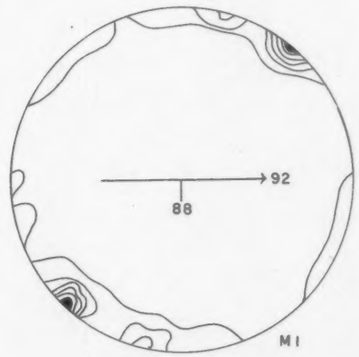

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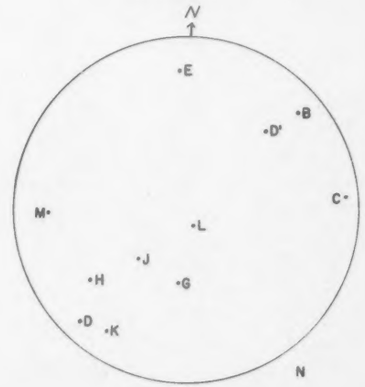

N

PETROFABRIC DIAGRAMS

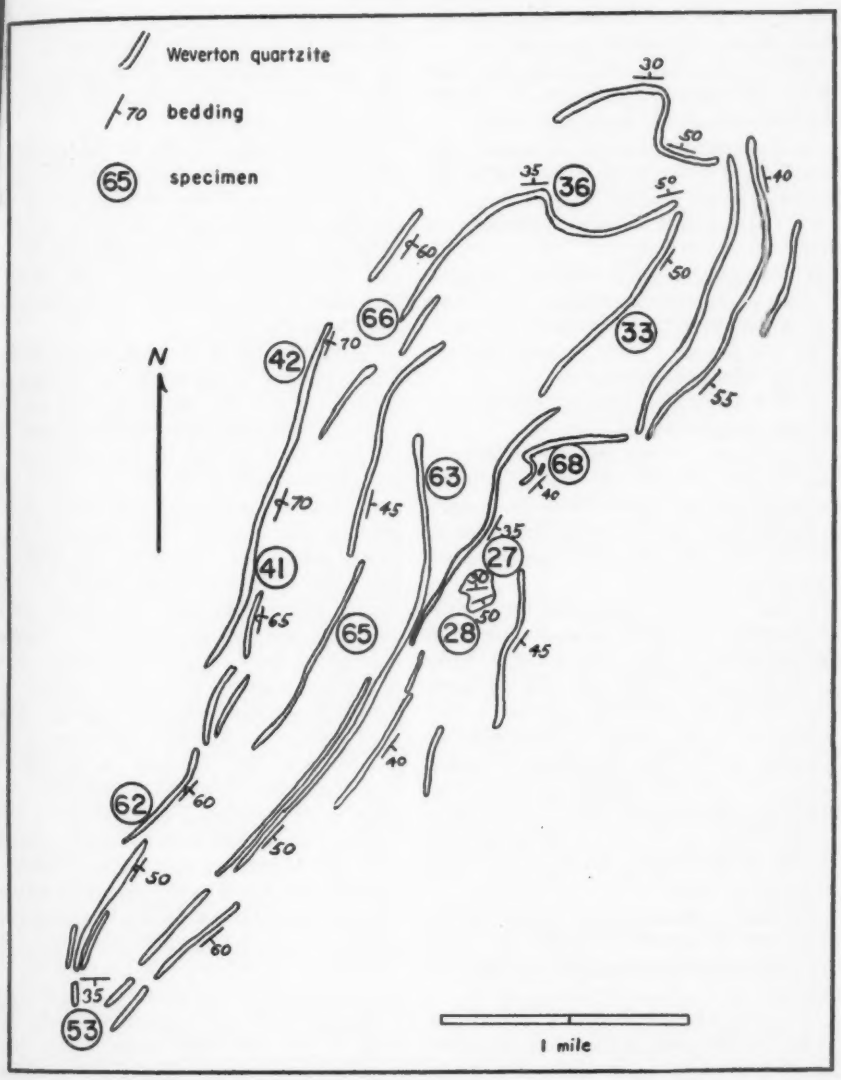

FIGURE 13.—SKETCH MAP OF SUGARLOAF MOUNTAIN ANTICLINE SHOWING PLANES OF LIQUID INCLUSIONS

is apparently at the zone where *ac* joints and the related planes of liquid inclusions coexist.

SUMMARY AND CONCLUSIONS

The following major conclusions were reached as a result of the study of the Sugarloaf Mountain area:

(1) Sugarloaf Mountain is an anticlinal dome.

(2) The local phyllites, formerly considered volcanics, may be correlated with the Lower Cambrian Harpers phyllite.

(3) The quartzites which constitute the mountain lie stratigraphically below the Har-

pers phyllite and are therefore most likely the Lower Cambrian Weverton quartzite.

(4) The local area, as well as the area as far west as Harpers Ferry, lies within the chlorite metamorphic zone and greenschist facies.

(5) The "Martic Overthrust" is possible, but local faulting seems to be a more likely explanation for the stratigraphic discontinuity.

(6) The Sugarloaf anticline is apparently a shear fold.

(7) Petrofabric analysis indicates a rotation of cleavage planes during folding from a large angle to bedding to their present orientation.

(8) The planes of liquid inclusions are apparently related to *ac* joints.

A summary of the evidence which leads to these conclusions follows:

(1) Probably the best evidence of the anticlinal nature of Sugarloaf Mountain lies in the orientation of bedding which boxes the compass in the northern and southern end of the anticline and dips outward except in the overturned western limb.

On the summit of a ridge at the north end of the mountain is an outcrop of quartzite in which the bedding dips to the northeast and the strike curves from 87° to 130° indicating the north nose of a northeast-plunging anticline.

Other evidence is found in the angular relationship of cleavage and bedding. On the east limb of the fold, the cleavage is steeper than bedding indicating the east limb of an anticline; on the west limb, the cleavage is less steep or parallel to the bedding indicating the west limb of an anticline.

The fold axes in the Harpers phyllite plunge away from the mountain opposite the north side, indicating that the large fold is a plunging anticline.

(2 and 3) On the west side of Frederick Valley, 8 miles west of Sugarloaf Mountain, the Harpers phyllite and Antietam quartzite lie in normal sequence above the Weverton quartzite on the anticlinal Catoctin Mountain (E. Cloos, oral communication). Under the anticlinal interpretation of Sugarloaf Mountain, the local phyllites lie above local quartzites just as the known Harpers phyllites lie above the known Weverton quartzite 8 miles to the west. Quartzitic layers common in the Harpers phyllite on South Mountain are also found in the phyllite of the Sugarloaf Mountain area. Since the local phyllite and quartzite are megascopically and microscopically almost identical with the known Harpers and Weverton and since their stratigraphic relationship is the same, a correlation seems assured.

(4) The Harpers phyllite in the Sugarloaf area and as far west as Harpers Ferry consists chiefly of quartz, sericite, chlorite, and albite. Porphyroblastic chlorite is found at Harpers Ferry but becomes more common toward the east. This mineralogy places the rock and the metamorphic terrane within the chlorite zone and greenschist facies.

(5) Faulting along the west border of the area is revealed by the contact of the Harpers phyllite and Frederick limestone which are not normally stratigraphically contiguous, and by the abnormal orientation of lineation and cleavage near the trace of the possible fault north and south of the known fault. However, no evidence was found of displacement of overthrust proportions. The same formations on either side of the possible fault trace are very similar lithologically and in grade of metamorphism. Although an overthrust is possible, a more likely explanation, involving expenditure of less kinetic energy, is a local thrust fault.

(6) Shearing and flowage were apparently the chief mechanisms involved in the formation of the Sugarloaf fold. The beds on the western limb of the fold are considerably thinner than those on the east limb (Fig. 10). Since the bedding and cleavage are frequently parallel on the west limb and always intersect at an angle on the east limb, it is postulated that shearing along the cleavage thinned the bedding on the western limb and thickened it on the eastern limb.

A drag fold on the east limb of the anticline indicates the inner beds move upward in relation to the outer beds. Such movement is possible in an anticline only under shear folding.

Fold axes in the Harpers phyllite opposite the north nose of the anticline change in orientation from parallel to the large fold to east-west. The explanation is believed to lie in the differential shear produced between the large quartzitic anticline and the incompetent phyllite during folding.

(7) The petrofabric study of mica and quartz suggests a rotation of the S_2 (flow cleavage) planes during deformation. Quartz maxima tend

to lie in the megascopic cleavage plane which is therefore considered to be the latest S_2 plane. Since mica maxima indicate the presence of earlier S_2 planes, the direction of rotation may be determined. The cleavage was apparently rotated during folding from an orientation at a large angle to bedding to its present position parallel or at a low angle to bedding. The rotation was not continuous but evidently proceeded in stages indicated by the various S_2 planes which lie between the oldest and youngest. However, the rotation probably occurred during just one major deformation.

(8) The geographic orientation of the planes of liquid inclusions in quartz found in the Weverton quartzite on Sugarloaf Mountain tend to parallel the prominent ac cross joints. The lack of abundant data, however, make the results less conclusive than is desirable.

It has not been a part of this investigation to study the structure and metamorphism of the area east of Sugarloaf Mountain. However, a continuous belt of phyllite exists between the Sugarloaf Mountain area and the Wissahickon schist of the eastern Piedmont. The age of this crystalline schist has long been in doubt. Further work will probably reveal a continuous gradation with increasing metamorphism of the Harpers phyllite of the Sugarloaf area into the Wissahickon schist lying about 12 miles to the east, at its westernmost extent, thus establishing the age of the Wissahickon as Lower Cambrian.

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Short Notes

INADVISABILITY OF RESTRICTING TERMS THAT DESIGNATE THE RELATIVE

BY J. E. EATON

Most terms used by man to label phenomena are necessarily pigeonholing ones. Groping among relations complex beyond comprehension, he pigeonholes (though not excluding change) as if a face were independent of other faces. Whatever the errors involved, the practice is mechanically necessary as regards routine conveying of conceptions. To average, to approximate, to speak as if a particular face observed is the whole, is more or less required in elemental communication.

However, perhaps since the earliest camp fires, and as recorded since at least the time of the ancient Greeks, man has realized, in varying degree, that faces or forms that he observes are related in complex ways not only to less obvious faces of a thing designated, but in infinitely complex ways, in time and space, to other faces in the universe. He currently has reasons to believe that even his so-called elements, given time and conditions, pass one into another; that they, too, are transient faces that are relative. Accordingly, though necessarily clinging to thousands of pigeonholing words for elementary communication, he has created a small handful of terms to designate ways of Nature, as differentiated from transient faces that he pigeonholes. The family name of this small handful of terms is the word *relative*. The family name, by itself, is universal. Two qualifiers are commonly required when we use it in connection with a particular phenomenon. Under and of the family name are various terms with various meanings, which can be qualified in various ways to compare phenomena, but which, whether qualified to refer to one phenomenon or used alone to imply infinity, designate the relative.

The gap between man's pigeonholing and the relative is unbridgeable. There can apparently be no term that conveys simultaneously the static pictures that man forms, and combines to conceive of change, and the kinetic relativity of Nature; if there were such a term no man, necessarily operating through his succession of pictures, could understand it. The most, seemingly, that we can do is to conceive a picture, and the next instant warn that it is a relative thing. That is, pigeonhole, and then immediately invoke the relative. There is no fusion, for the two classes are irreconcilable, but first a conceiving of one, and then a sensing of the other.

There are a few terms—rare, invaluable—which have been evolved by man to warn himself when he pigeonholes. These are nouns which in meaning designate the relative, but which have the peculiarly useful trait that they can be substituted, by adding to them a single qualifier, for any pigeonholing word, and, when deliberately substituted for or added to the latter, warn that the thing so referred to is relative to other things. When we deliberately label by one of these terms a picture that we must use, we see the picture, and then we sense the relative. We realize that the particular picture is a means to convey the latter. That, though we ostensibly refer to the picture, what is actually designated is the relative.

Chief among terms having the particular significance is the word *facies* (Latin: face, form). To what extent the ancient Romans used the word to include a meaning more than to name pictures conceived by man is uncertain. But use of the word in modern science has been in the extended sense: that a thing so designated is referred to as being only a face, a form. Thus, while the words face and form are variously used in either a pigeonholing or a relative sense, the term facies, when deliberately used, designates only the latter. The term facies applies to every thing in the universe, from the largest to the smallest; there are reasons to believe that no thing escapes being a face, a form. The word has little use in elementary communication,

where we accept a picture as if this were the whole. In science it immediately acquires large significance. As we progress in this it increases in importance, until perhaps no term has more significance, except the family name.

In geology we utilize many valuable pigeonholing words; for example: sand, shale, formation, species, genus. Such are very useful; we could not convey impressions without them or an equivalent. When we use them we convey some picture, as diverse as the number of viewers, and as the number of hearers or readers. The result is approximations by giver and taker, none of whose pictures exactly match, and are sometimes quite different. Nevertheless, the practice is commonly useful and even necessary. A picture by which we pigeonhole tends to be conceived, by us and others, as some kind of a unit; qualifications may perturb us, but these tend to be secondary to a conceived unity. If we substitute the term facies for, or add it to, any pigeonholing word, we reverse the emphasis. We refer to the picture, as before, but now this becomes secondary to qualifications; it is now a vehicle to convey something else. We see the thing ostensibly referred to, but it is no longer separate and alone. Beyond it, we sense the inconceivable complexity of the relative. We realize that this, not the vehicle, is the meaning actually designated. As: "(This sand is) A facies." . . . a face, a form in Nature. In effect, when we use the term facies for or with the name of a thing that we pigeonhole, a red light glows.

Words which refer solely to the relative, and those which refer solely to some pigeonhole, tend to be safely in their respective meanings; the difference between the two irreconcilable classes is too great for the meanings to be successfully attacked. The invaluable terms of dual purpose, which refer to a thing pigeonholed but in meaning designate the relative, are, on the other hand, subject to continual attack. We are so constituted that we endeavor to assign each picture that we conceive to some niche. We accept the infinitely relative on faith, as a misty thing beyond comprehension. But an individual picture, we feel, should be in one class or the other. A term that mentions both (at once) tends to affect us somewhat as does a double exposure on a photographic film. Some persons can resolve the two images in alternate seconds; others can see an image only by concentrating on one. The result is continual conflict as regards the rare terms of dual purpose: defense by those who can resolve the images, and attack by those who can see but one. Due to pressure from the more literal-minded, the tendency is toward reducing the invaluable terms of dual purpose to the status of pigeonholing words.

As regards the term facies, it is apparent, from their applications, that some geologic workers are unaware of its derivation and regard the term as a recent concoction by geologists that applies solely to geology and is its private property; as a current fad; as being but a generalized alternate noun. They use the word as no more than an alternate for certain pigeonholing words, and as having no more meaning than these. They do not see the red light. Then there are workers who catalogue each conceived phenomenon according to some ideology expressed in some straight-line chart. Not in the tentative, approximating way that we commonly make a cast, but as a particular system. Each of numerous conceived things in inexact science is chopped off or stretched by them to fit some niche. And each niche is designated by some one of bewilderingly numerous or complex names, which the particular ideologist has spent years memorizing, but which to the rest of us may cause increased confusion. When we reflect on the countless competing ideologies of man, each with its particular choppings-off, stretchings, and names to be memorized, it is apparent that the human mind is not capable of digesting all, or even an appreciable part, of these. (Many ideologists seem not to realize, in their sincere purpose of clarifying things, that by making niches and names sufficiently numerous and complex they can defeat the very purpose that they would effect).[1] The ideologist

[1] In some of the so-called exact sciences such as mathematics and chemistry, where we use essentially uniform building blocks, groupings may profitably be given a great number of accretionary names which reflect self-explanatory aggregates. In so-called inexact sciences, on the other hand, such as medicine, psychology, and geology, the worker chooses from a chaos of building blocks having every conceivable irregularity. He has, of course, a proportion of somewhat uniform blocks borrowed from the exact sciences as isolated bases. But at his first step, and every subsequent one, he must try to fit the irregular blocks, having dimensions ofttimes so

has scarcely finished his particular system when along comes the term facies and warns him that each thing on which he has based his convictions is only a face, a form. Between those who do not see the red light and those who are irked by its warning, there is continual pressure toward reducing the term facies to a pigeonholing status.

Some have proposed changing the meaning of the word facies from the classic definition (face, form) that covers every thing in Nature and whose essence is the relative. It has been proposed that the meaning of aspect be used, which would merge the term facies with the thousands of nouns that designate aspect. But facies means more than aspect; indeed, it cautions against this. As commonly used it implies relations in time and space—the relative, whereas aspect means appearance more or less apart from such. Face, form, and aspect, considered individually, are all used in either a pigeonholing or a relative sense. Value and appreciation of the term facies in modern science has depended on the feature that, as usually adopted, it implies only the relative. Its value lies in reminding us of the ways of Nature, not those of man. Our thousands of pigeonholing nouns designate aspect. The term facies warns against aspect.

The term is not restricted as to kind, size, propinquity, succession, or time; though, like all terms that designate the relative, it can be qualified to bear on any of these in a given instance. It applies quite as well to an unknown atom, or nebula, beyond the most powerful telescope as it does to any face known to man. It is not available as the fixed name for a kind, size, whole or part, under any classification, though it may be applied, to indicate relativity, to any conceived whole or part from the smallest to the largest. Like the family name, it is universal. In essence, it is a static form of the term relative by which we grasp at the latter; as we might slow a stream of cinema film, call attention to one or more frames, and imply that these are part of a continuing, kinetically operating whole. The term facies treats the continuous, ungraspable stream of the relative as if this were broken into parts. By its use, man, conceiving through his static pictures, views the unbroken, unpausing, ungraspably continuous stream in pieces that he might grasp. His mechanics, as mentioned, are to pigeonhole a thing as if this were separate, and then, by calling it a facies, remind himself that it is related to other things in space and time; that it is relative.

Some, viewing the word facies as if best applicable to things of certain sizes, kinds, proximities, or other particular nature, have favored restricting it to such. The root meaning, universal in application, does not lend itself to such restriction. The word may be inflected or qualified as one pleases, but its base meaning is ancient and clear. For several thousand years it has meant face, form.

Man necessarily operates through a succession of pictures. If these involve sameness of impressions he conceives a thing to be static. If they involve differing impressions he conceives of change. If they change too rapidly for him to separate (*cf.* cinema, television) he conceives of motion. When he perceives or analyzes he does so through static pictures, whether noted as such or speed-jumbled to convey motion. Nearly all of his common nouns designate pigeonholings by him; a system of niches more or less separated. He easily conceives of one niche changing with time or place to another; he sees profuse examples. To a limited extent he conceives of simultaneous relative bearings between two or more niches, but, other than direct, simple sense comparisons, these tend to be shadowy. Beyond direct, simple sense comparisons he relies on symbols that outline abstractions; and these, like all of his pictures, inevitably fall into niches. Because of his make-up, man's niches are commonly not only useful, but necessary.

However, man has for long been aware that his succession of static pictures is not the way of Nature, but that she acts without pause. There are reasons to believe that whether she moves with the speed of light, or with that of change in the most nearly stable element, she

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dim that he can only feel for and guess at these, with the more uniform ones and with each other. The process is more one of personal judgment, casting through trial and error, than of mechanical combination. In such a process, precise niches and a complex system of names for these may pyramid error rather than clarify.

acts as an integrated kinetic whole whose parts are infinitely relative, each to all others. We are too poorly equipped, with our system of graduations, to assimilate such. We contact the relative, through our niches, by perceiving that some of these change toward others; a series of intermittent perceptions not unlike the niches, and dependent on them, that we make stand for the continuous stream we cannot grasp.

Man has evolved a few terms to warn himself when he pigeonholes. Chief among these is the word facies. It warns us that every thing in the universe, including those from which we gain our impressions, and on which we base our convictions, is only a face, a form. Insofar as man's recognition of his errors with respect to Nature, and his endeavor to decrease these, are concerned, it is the first executive term of the relative.

The word facies, universal in application, is not solely the property of the science of geology. It is a heritage of the entire race. If we in geology restrict it, we may feel sure that other of the sciences will restore its meaning.

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